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(Continued on inside back cover)

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The printing of this publication has been approved by the Director of the Bureau of the Budget, February 11, 1952

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A GENERALIZED STUDY OF PRECIPITATION FORECASTING

PART 1: COMPUTATION OF PRECIPITATION FROM THE FIELDS OF MOISTURE AND WIND

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ABSTRACT

For the eventual purpose of developing a generalized method for predicting precipitation from assumed accurate prognoses of the required meteorological elements, a preliminary investigation is made of the contemporary relationship between precipitation and the fields of moisture and wind in the atmosphere. Procedures are developed for calculating the rate of precipitation by computing the divergence of the horizontal wind at 50-mb. intervals from the surface to 300 mb. using a technique suggested by Bellamy. From these values, vertical speeds are determined. Using the vertical speeds in Fulks' formula for the rate of precipitation from pseudo-adiabatically ascending air, but with some modifications to compensate for non-saturated initial conditions, a method is derived for calculating the intensity of precipitation. Computed amounts are compared with observed precipitation.

INTRODUCTION

The calculation of precipitation intensity from a knowledge of other atmospheric characteristics is a subject which, in addition to possessing features of purely academic or scientific value, has become of increasing practical interest during the past few years. In part this interest arises because of a gradual increase in the number and accuracy of meteorological observations, which makes feasible the use of more precise methods of calculation; in part it stems from a greater need for quantitative information on precipitation intensities for hydrometeorological work, cloud seeding evaluation, etc.; and in part it stems from the development of new methods for predicting some of the atmospheric variables which are related to precipitation occurrence and intensity as, for example, the use of modern electronic computing equipment to provide a prediction of the height contours at 500 mb. and thus also a prognosis of the "gradient" wind field at that level [1].

The present study was started primarily in an attempt to use these latter developments as a means of forecasting precipitation, under the assumptions (*a*) that the contour fields on constant pressure surfaces will be predicted with increasing efficiency and accuracy as studies such as those making use of electronic computers proceed, (*b*) that other meteorological variables, such as the field of moisture, will be subjected to similar treatment, and (*c*) that it will then be necessary to provide methods for calculating, from these predicted variables, the weather elements about which forecasters and the public in general are primarily concerned—in this case, the occurrence and amount of rainfall. It may be noted in passing that these assumptions are in turn predicated on the hypothesis that it is essential, or at least desirable, that the prediction process should proceed in this somewhat indirect fashion, i. e., that prognostic charts of certain physical quantities, of interest primarily to the meteorologist, should first be provided and that from these data the weather forecast should be made. Whether or not this is the most logical or efficient procedure, or currently provides the most accurate weather prediction, are questions of the basic philosophy of weather forecasting which are beyond the province of this paper to discuss. It is considered sufficient to note here that the construction of such prognostic charts, either mental or formal, is a part of most forecasting procedures at present and probably will continue to be so for some time to come.

The general problem being studied may accordingly be

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defined: To provide a procedure for computing the amount of precipitation to be expected, having been given a prognosis of those quantities which are related to the precipitation process. As will be apparent from the succeeding discussion, this paper can be considered as only a preliminary report on the progress of work during the first stages of the investigation; plans for continuing the study are discussed briefly at the end of the paper.

METHOD OF ATTACK

In attacking this problem, two types of approach are possible—the statistical and the physical. The statistical approach, in which variables, chosen because synoptic experience and/or physical reasoning indicate that they are related to the element being forecast, are combined graphically or in regression equations, has been used in studies by Carstensen and Hardy [2], Bristor [3], and others.[2] Such procedures are unquestionably useful when the basic relationships between predictor variables and the weather element being predicted are not known quantitatively or when, because of difficulties encountered in solving mathematical expressions, the fundamental equations cannot be used in a practical way. In the approach being used here, however, it is desired to start with whatever basic knowledge is presently available and, by making such assumptions and approximations as are necessary to solve the resulting equations, to build up a generalized relationship between the variables obtained from prognostic charts and the concurrent precipitation. The procedure should therefore be independent of geographic location, season, and similar restrictions which usually apply to statistically derived methods; or if it has initially certain weather or geographical limitations, at least the assumptions which make the restrictions necessary will be known.

In approaching the problem it was considered desirable to limit the study to the computation of precipitation associated with large-scale atmospheric disturbances, i. e., middle latitude cyclones. Such storms are productive of the majority of widespread heavy precipitation situations and are accordingly of considerable importance from both the scientific and economic standpoints. They are also better suited to the type of analysis used here than are small-scale disturbances of the size of single thunderstorms. It was further considered desirable to begin with the fewest assumptions possible, and to assess in turn the validity of each modification which would be required in order to develop an operationally practicable forecasting procedure. Thus, in this study, "true" winds from RAWIN observations are used as a measure of the velocity field, leaving for further investigation an evaluation of the validity of the "gradient" wind and other similar approximations which may be necessary in order to provide a practical forecasting method. It should be noted, however, that some of these latter devices might be expected to filter out some of the small-scale eddy effects which are unavoidably present in RAWIN measurements, and thus may actually provide better approximations of the large-scale precipitation processes than the winds used in this study. Further discussion on this point is given at the end of the paper. Since it has been assumed that the wind and moisture fields will be forecast with sufficient accuracy and sufficiently far in advance to provide reliable predictors for determining the weather to be expected, only the contemporary relationships between these fields and the intensity of precipitation will be considered here.

BASIC ASSUMPTIONS

If it is assumed that cooling processes other than those associated with adiabatic lifting are small, that sufficient and suitable nuclei are present, and that ideal pseudo-adiabatic conditions exist, then the intensity of precipitation may be determined from the fields of moisture and vertical component of velocity. Considering the first of these assumptions, Möller [4] has shown that cooling by radiation from clouds which have an "average distribution" in a low pressure area in middle latitudes is, in the mean, less than 2° C. per day for clouds at altitudes below 25,000 feet. The radiational cooling may be as great as 15° C. per day in the advance portion of a cyclone where the cloud exists as a thin sheet at 25,000–30,000 feet, but thin clouds at this altitude rarely produce precipitation at the surface. For cloud masses with bases below 800 mb. and whose thickness exceeds 200 mb., Hoffer [5] indicates that the rate of cooling is less than 5° C. per day. The rate of cooling due to moderate pseudo-adiabatic lifting, on the other hand, is about 25° C. per day and may exceed 60° C. per day for vertical speeds of 10 cm sec.$^{-1}$ or more. Accordingly, while the effect of radiational cooling is not completely negligible, it may as a first approximation be considered small in comparison with cooling due to ascending motion, especially in those cases where cloudiness is thick enough and exists at sufficiently low levels to produce appreciable precipitation.

Other nonadiabatic temperature changes, i. e., those due to heat exchange between the cloud elements and the air, and heat conduction in the cloud interior, undoubtedly produce structural changes in individual clouds or thunderstorms, but for large-scale processes of the order of a cyclonic disturbance, the effects are unimportant.

The assumption that a sufficient number of nuclei are always present is one which, because of incomplete knowledge of the basic mechanisms of condensation and precipitation in the atmosphere, apparently cannot be adequately evaluated at present. Furthermore, lack of regular observations of the nature and number of atmospheric nuclei which could be used in any systematic way in day-to-day calculation of precipitation, eliminates consideration of this effect in any practical way.

Concerning the final assumption, Möller [6] states that although a "general" pseudo-adiabatic process (i. e., one in which part of the condensed moisture drops out as

[2] See adjoining article by Jorgensen.

precipitation and part is carried along in the ascending air) is the normal situation in nature, this process "cannot theoretically be evaluated since no numerical estimate can be made of the amount of precipitation elements which drop out or are carried along." Accordingly, it has been necessary to neglect, in this preliminary study, the problem of moisture storage within the cloud. For a similar reason the evaporation from falling rain has been neglected and the assumption made that all of the water condensed by pseudo-adiabatic cooling is realized in the form of precipitation.

RATE OF PRECIPITATION

The rate of precipitation from a thin layer of saturated air ascending pseudo-adiabatically has been derived by Fulks [7] by assuming that the change in thickness of the rising air may be neglected and by using an approximation for the change in vapor pressure with height in one place in the derivation. The errors which result from either or both of these devices are shown by Fulks to be small in comparison with other factors which affect the pseudo-adiabatic process. The rate of precipitation r is then given by:

$$r=\left[\frac{\epsilon}{RT}\left(\frac{de}{dz}+\frac{eg}{RT}\right)\right] V_z\Delta z=IV_z\Delta z \tag{1}$$

where ϵ is the ratio of the density of water vapor to that of dry air at constant pressure and temperature
R is the gas constant for dry air
T is the mean absolute temperature of the layer
e is the saturation vapor pressure of the layer
z is height
g is the acceleration due to gravity
V_z is the vertical component of velocity of the layer, positive upward
Δz is the thickness of the layer

The quantity I, representing the expression enclosed within the brackets, is the rate of precipitation per unit of vertical speed and thickness of the layer; it is introduced here in order to simplify the notation in the section which follows. Since, for saturated conditions, the vapor pressure is determined by the temperature of the layer and de/dz is determined by its temperature and pressure, the rate of precipitation can be obtained from a knowledge of the temperature, pressure, vertical velocity, and thickness of any shallow layer. Accordingly, if the atmosphere is divided into a number of such layers and the contribution r_i of each layer is computed, the total rate of precipitation P from the column will be,

$$P=\sum_i r_i \tag{2}$$

where index i ranges from the surface to the top of the atmosphere.

Figure 1, adapted from Fulks' paper [7], is a graphical solution of equation (1) for the normal range of atmospheric pressures and temperatures. It provides a method for computing the rate of precipitation from an air column at any instant if the air is assumed to be saturated initially, and the vertical velocity, temperature, pressure, and thickness of each layer in the column are known. Since these latter quantities can be regularly obtained only at finite time intervals in the atmosphere, the convention is adopted in this study that the computed precipitation rates will be used to calculate the amount of precipitation during the 12-hour period between RAOB-RAWIN measurements. Using this device, it is possible to modify Fulks' formula so as to account for, and to some extent eliminate, the assumption of initially saturated conditions.

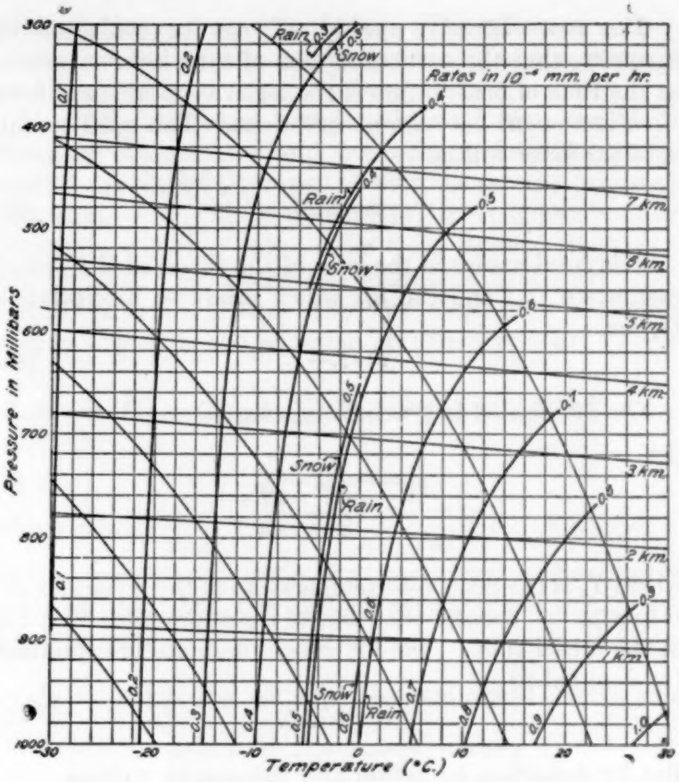

FIGURE 1.—Rates of precipitation from adiabatically ascending air for a one-meter layer with a vertical velocity of one cm. per sec. (After Fulks [7]).

MODIFICATION FOR NON-SATURATION

The total amount of precipitation, M, contributed by a thin layer during the time, Δt, is

$$M=r\Delta t'=IV_z\Delta z\Delta t' \tag{3}$$

where $\Delta t'$ is the portion of Δt during which the air is saturated. While $\Delta t'$ can be computed from the vertical velocity and moisture available, it is more convenient for computing purposes to change equation (3) to the form

$$M=IV_z'\Delta z\Delta t, \tag{3a}$$

where V_z' is defined by the equation

$$V_z'\Delta t=V_z\Delta t'. \tag{3b}$$

This new "effective speed," V'_z, can be evaluated by assuming that the vertical motion of each layer measured at the time of observation is the mean vertical speed for a time increment between observations. The total ascent, z, of the layer will be

$$z = V_z \Delta t. \tag{4}$$

This total ascent is the sum of the lift needed to reach saturation (z_1) and the lift which produces precipitation (z_2) or,

$$z = z_1 + z_2. \tag{4a}$$

The lift needed to reach saturation is

$$z_1 = -\frac{T_o - T_{do}}{\frac{dT}{dz} - \frac{dT_d}{dz}} \tag{5}$$

where T_d is the dew point temperature of the layer and the subscript "o" denotes the value of the appropriate variable at the initial time. The lift which produces precipitation (z_2) is equal to

$$z_2 = V_z \Delta t'. \tag{6}$$

But by definition (equation 3b), this can be written

$$z_2 = V'_z \Delta t \tag{6a}$$

and the quantity V'_z can be evaluated by combining equations (4) through (6a) giving,

$$V'_z = V_z + \frac{T_o - T_{do}}{\Delta t\left(\frac{dT}{dz} - \frac{dT_d}{dz}\right)}. \tag{7}$$

Substituting in this equation $\Delta t = 12$ hours, $\frac{dT}{dz} = -9.8°$ C. km.$^{-1}$, and $\frac{dT_d}{dz} = -1.6°$ C. km.$^{-1}$ gives for vertical speeds in cm. sec.$^{-1}$,

$$V'_z = V_z - 0.28(T_o - T_{do}). \tag{7a}$$

Introducing these approximations for the adiabatic lapse rate of temperature and the lapse rate of dew point with height introduces only negligible errors in the computations.

VERTICAL VELOCITY

An essential part of the problem is concerned with methods of computing the vertical velocity. Because of its importance in many meteorological processes, a number of investigators have studied procedures for computing vertical motion; and although it would be impractical to discuss the work of all who have studied the problem, it is desired to note that several recent papers have been particularly concerned with attempts to develop relationships between precipitation and vertical motion computed in various ways. The association of vertical motion and precipitation has been studied by Byers and Rodebush [8] who computed the horizontal divergence due to the sea breeze in the Florida Peninsula and showed its relationship to the occurrence of thunderstorms, Miller [9] who, under the assumption of adiabatic motion, related large scale vertical motions to precipitation, Das [10] who has suggested the use of divergence charts for the prediction of precipitation in India, and Bannon [11] who reversed these ideas and obtained an estimate of vertical motion from the rate of rainfall.

In this study, the vertical velocity was obtained by using a method suggested by Bellamy [12]. In this procedure, the horizontal divergence within a traingular area enclosed by three RAWIN (or PIBAL) stations is computed by assuming a linear wind field between points; these results are then used in an equation of continuity, neglecting local changes and horizontal variations in density, to give the vertical velocity at the top of any layer:

$$V_{z2} = \frac{\rho_1}{\rho_2} V_{z1} - \frac{1}{2}\left[\frac{\rho_1}{\rho_2} D_1 + D_2\right] \Delta z. \tag{8}$$

Here the subscripts "1" and "2" refer to the bottom and top of the layer, respectively, ρ is the density of the air, and D is the horizontal divergence. Neglecting the local change and horizontal variations in density will result in errors of less than two percent in the vertical velocity. Other errors arise because of (a) the assumption of linearity in the wind field and (b) the inaccuracy and nonrepresentativeness in the observed winds. It should be noted, however, that the assumption of linearity is not completely necessary since the validity of the computation also holds if the mean value of the wind component normal to a side of the triangle is equal to the average of the values at the end points. Variations in the wind of a spatial scale smaller than that of the triangle may be neglected here, since only large scale processes associated with middle-latitude cyclones and anticyclones are being considered.

Errors due to inaccuracies in the wind observations are more important, and are probably of the same order of magnitude as the divergence. However, in the procedure suggested by Bellamy, the vertical velocities are obtained by a process of first summing a group of three partial divergences horizontally and then adding these horizontal divergences through a vertical column. This means that the vertical motion at the top of each layer, and consequently the precipitation contributed by the layer, is determined by at least six wind observations, with the total number of observations used increasing directly as the number of layers used in the computations. Accordingly, if the errors in the observed winds are random, it is possible for the sum of the errors in the computed vertical motions to converge toward zero. In an attempt to check on this possibility, a random difference, i. e., "error" was introduced into the wind direction by using winds smoothed to the nearest of the 16 principal points of the

compass. The vertical motions, by layers, computed from these "smoothed" winds showed but little deviation from the vertical velocities obtained using the winds calculated to the nearest degree, thus indicating that these random differences in the winds did indeed converge during the summing process. While it appears reasonable to assume that the actual observational errors in the winds used in the horizontal sums are random, it is probable, on the other hand, that those in the vertical summations are to some extent correlated. Consequently, it is not possible to determine *a priori* whether or not the errors in the computed vertical velocities (and thus the precipitation) will also be small. This must be determined by an examination of the errors in computed precipitation; further discussion of the matter will therefore be deferred until later in the paper.

By noting that $V_z=0$ for a level ground surface, the vertical speed at the top of the first layer, from equation (8), may be written:

$$V_{z1}=-\frac{1}{2}\left[\frac{\rho_o}{\rho_1}D_o+D_1\right](z_1-z_o). \tag{9}$$

Here, and in what follows, the subscript "o" refers to the surface, "1" to the top of the first layer, etc. Substituting this expression for V_{z1} in equation (8) and rearranging and collecting terms gives,

$$V_{z2}=-\frac{1}{2}\frac{\rho_o}{\rho_2}D_o(z_1-z_o)-\frac{1}{2}D_2(z_2-z_1)-\frac{1}{2}\frac{\rho_1}{\rho_2}D_1(z_2-z_o). \tag{10}$$

If this process is continued, layer by layer, the general expression for the vertical speed at the top of the i th layer will be seen to be given by,

$$V_{zi}=-\frac{1}{2\rho_i}\left[\rho_o(z_1-z_o)D_o-\rho_i(z_{i+1}-z_i)D_i+\sum_{j=1}^{i}\rho_j(z_{j+1}-z_{j-1})D_j\right]. \tag{11}$$

For the purposes of practical computing, layers 50 mb. in thickness are selected as being sufficiently representative of the vertical distribution of both the wind and moisture fields. It is then assumed that the surface is at 1,000 mb., that the density variation in the vertical is that of the U. S. Standard Atmosphere, and that the divergence at 950 mb. is representative of the layer which extends from the surface to that height. The first of these approximations may produce an extreme error in the computed precipitation of the order of 0.05 inch per 12 hours where the pressure is 25 mb. higher or lower than the assumed value; if the difference exceeds 25 mb., the integration may of course be started at the nearest 50-mb. level. The second assumption introduces a negligible error (less than 1 percent) in the computed precipitation. Use of the last device minimizes undesirable surface turbulence effects which it is desired to eliminate in computing large-scale characteristics of the circulation. The effect of surface friction on the large-scale divergence patterns is assumed small and has been neglected here.

Using the values of ρ and z obtained from the U. S. Standard Atmosphere and performing the summations over 50-mb. intervals, revealed that the terms $\rho_i(z_{i+1}-z_i)$ and $\rho_j(z_{j+1}-z_{j-1})$ in equation (11) were approximately constants, differing only in the third decimal place as i and j varied. Since only two significant figures are justified in the computations, these two terms may be considered as constants and equation (11) then written:

$$V_{zi}=-a_i(D_{950}-D_i)-b_i\sum_{j=950}^{i}D_j,\ j=950,\ 900,\ 850,\ \ldots,\ i \tag{12}$$

where $a_i=\frac{\rho_i(z_{i+1}-z_i)}{2\rho_i}$ and $b_i=\frac{\rho_j(z_{j+1}-z_{j-1})}{2\rho_i}$. Values of a_i and b_i, when the divergence at each level is obtained in units of sec.$^{-1}$ and the vertical speed in cm. sec.$^{-1}$, are given in table 1.

TABLE 1.—*Values of a and b for computing vertical velocities from equation (12)*

Top of layer (mb.)	a (cm.)	b (cm.)
300	0.58×10^{5}	1.12×10^{5}
350	.51	.99
400	.45	.89
450	.41	.81
500	.38	.74
550	.35	.68
600	.33	.64
650	.31	.60
700	.29	.56
750	.27	.53
800	.26	.51
850	.25	.48
900	.24	.47
950	.22	.44

PRACTICAL COMPUTING PROCEDURE

Equations (2), (7a), and (12) provide a method for computing the total amount of precipitation during a 12-hour period, having been given the fields of wind and moisture, and subject to the limitations imposed by the assumptions used in the derivation. Since, as was noted earlier, it is not possible to assess the magnitude of all of the errors introduced by these assumptions, and consequently to evaluate their total effect, it is necessary to apply the procedure to the atmosphere and to examine the data thus obtained. For the purpose of this evaluation, a triangular area in the midwestern United States with vertices at the RAOB-RAWIN stations at Columbia, Mo., Little Rock, Ark., and Nashville, Tenn. was selected. (See fig. 2.) This region is characterized by relatively flat terrain so that the disturbing effects of topography upon the large-scale vertical motions are minimized; it is also located close to the normal tracks of middle-latitude cyclones during the winter season, thus assuring that the observed precipitation would, for the most part, be due to larger scale atmospheric processes; finally, reasonably good data on

FIGURE 2.—Location of the triangular area selected for precipitation computations. The 15 stations in the interior of the triangle were used for verifying the computations.

winds, temperature, and moisture aloft, as well as surface precipitation were available for the area.

The results were verified by using an average of the precipitation recorded at 15 stations within the triangle, with the "mean" value being weighted by selecting more stations near the center than near the edges. (See fig. 2.) This areal distribution was chosen first, because "average" values of divergence and moisture were used in making the computations and the most likely place for the occurrence of these values (and consequently the resulting precipitation) would be near the center of the triangle; and second, because the errors introduced by neglecting the advection of moisture and divergence during the 12-hour period should, on the whole, be smaller near the center of the triangle, if it is assumed that this advection may occur from any direction.

Precipitation rates for an initial sample of one winter month were computed for each 50-mb. layer from the surface to 300 mb. and the divergence was computed from RAWIN data calculated to the nearest degree and in meters per second. Later, however, an examination of the data showed that, in the area selected, the contribution of the layers above 500 mb. during the winter months was usually small and the termination of most RAWIN observations at or below 500 mb. during rain or snow situations frequently necessitated dropping the computations at that level. Furthermore, as was pointed out earlier, the use of wind directions to the nearest 16 compass points (which are easily available from punched-card summaries) introduced no appreciable additional error in the computed precipitation. Consequently, an additional three months of computations were made for the same area using data only to 500 mb. and wind directions smoothed to 16 points of the compass.

An example of a complete computation is shown in tables 2 and 3. In table 2 the divergence, obtained from the winds by using a nomogram adapted from Bellamy [12], is entered in the first column as convergence (convergence=minus divergence) in order to minimize calculations with negative quantities. The last column provides the desired vertical velocities, the entire process representing a solution of equation (12).

If examination of the last column shows negative (downward) vertical speeds at all levels, no further calculations are necessary since the resulting adiabatic processes would tend to inhibit the cooling necessary for precipitation. If, on the other hand, the vertical speed is positive for any layer, it is necessary to see whether or not it is sufficiently large so that, in the 12-hour period between observations, the air will become saturated. This may be done by calculation from equation (7a), the vertical

TABLE 2.—*Example of procedure for computing vertical velocities. Convergence values (C) in the first column are obtained from the divergence (D) of the winds by defining $C=-D$. The last column is the vertical velocity. Data are from the 1500 GMT observations of Dec. 5, 1950*

Pressure (mb.)	C (sec.$^{-1}$)	$C_{950}-C_i$ (sec.$^{-1}$)	$\sum_{j=950}^{i} C_j$ (sec.$^{-1}$)	$a_i(C_{950}-C_i)$ (cm. sec.$^{-1}$)	$b_i \sum_{j=950}^{i} C_j$ (cm. sec.$^{-1}$)	V_{si}: $a_i(C_{950}-C_i)+b_i \sum_{j=950}^{i} C_j$ (cm. sec.$^{-1}$)
300	Missing					
350	11.0×10^{-6}	−11.8×10^{-6}	36.5×10^{-6}	−6.5	36.4	29.9
400	2.4	−4.2	25.5	−1.9	22.7	20.8
450	2.5	−4.3	23.1	−1.8	18.7	16.9
500	3.7	−5.5	20.6	−2.1	15.2	13.1
550	2.5	−4.3	16.9	−1.5	11.5	10.0
600	3.3	−5.1	14.4	−1.7	9.2	7.5
650	2.7	−4.5	11.1	−1.4	6.7	5.3
700	2.8	−4.6	8.4	−1.3	4.7	3.4
750	1.7	−3.5	5.6	−.9	3.0	2.1
800	2.0	−3.8	3.9	−1.0	2.0	1.0
850	3.1	−4.9	1.9	−1.2	.9	−.3
900	.6	−2.4	−1.2	−.6	−.6	−1.2
950	−1.8	0	−1.8	0	−.8	−.8

TABLE 3.—*Example of procedure for computing total precipitation. Values of V_s in the first column are obtained from table 2. Data are from the 1500 GMT observations of Dec. 5, 1950*

Pressure (mb.)	V_s (cm. sec.$^{-1}$)	$0.28(T_o-T_{do})$ (cm. sec.$^{-1}$)	V_s': $V_s-0.28(T_o-T_{do})$ (cm. sec.$^{-1}$)	r' from fig. 1 (mm.hr.$^{-1}$)	Δz (m.)	r: $V_s'r'\Delta z$ (mm. hr.$^{-1}$)
300	Missing					
350	29.9	3.9	26.0	0.01×10^{-4}	990	0.026
400	20.8	3.9	16.9	.01	880	.015
450	16.9	2.8	14.1	.04	805	.045
500	13.1	2.2	10.9	.10	740	.081
550	10.0	1.1	8.9	.18	685	.110
600	7.5	.6	6.9	.29	635	.127
650	5.3	.6	4.7	.35	595	.098
700	3.4	.6	2.8	.40	560	.063
750	2.1	2.0	.1	.37	535	.002
800	1.0	1.1	−.1			
850	−.3					
900	−1.2					
950	−.8					

$\Sigma r=P=0.567$ (mm. hr.$^{-1}$
$P=.27$ (in. 12-hr.$^{-1}$
Observed Precipitation $P_{obs}=.15$ (in. 12-hr.$^{-1}$

velocities at each level being obtained from table 2, and the values of T_o and T_{d_o} by averaging the values reported by the RAOB ascents at the vertices of the triangular area selected. If V'_z is negative for all layers, no further steps are necessary since it is presumed that the air will not be lifted sufficiently during the 12-hour period to produce precipitation. However, if V'_z is positive for any layer, it may be substituted for V_z in equation (1) and used to obtain the amount of precipitation contributed by that layer.

In this study, equation (1) was evaluated by making use of the nomogram, figure 1. Precipitation rates were then obtained by making the working assumptions that the initial dew point of the ascending air would be its condensation temperature, and that the initial pressure of the layer would be its condensation pressure. If the air is close to saturation, these assumptions introduce negligible errors, and if the air is far from saturation there will usually be insufficient lift to produce saturation and no precipitation will be computed in any case. However, for high

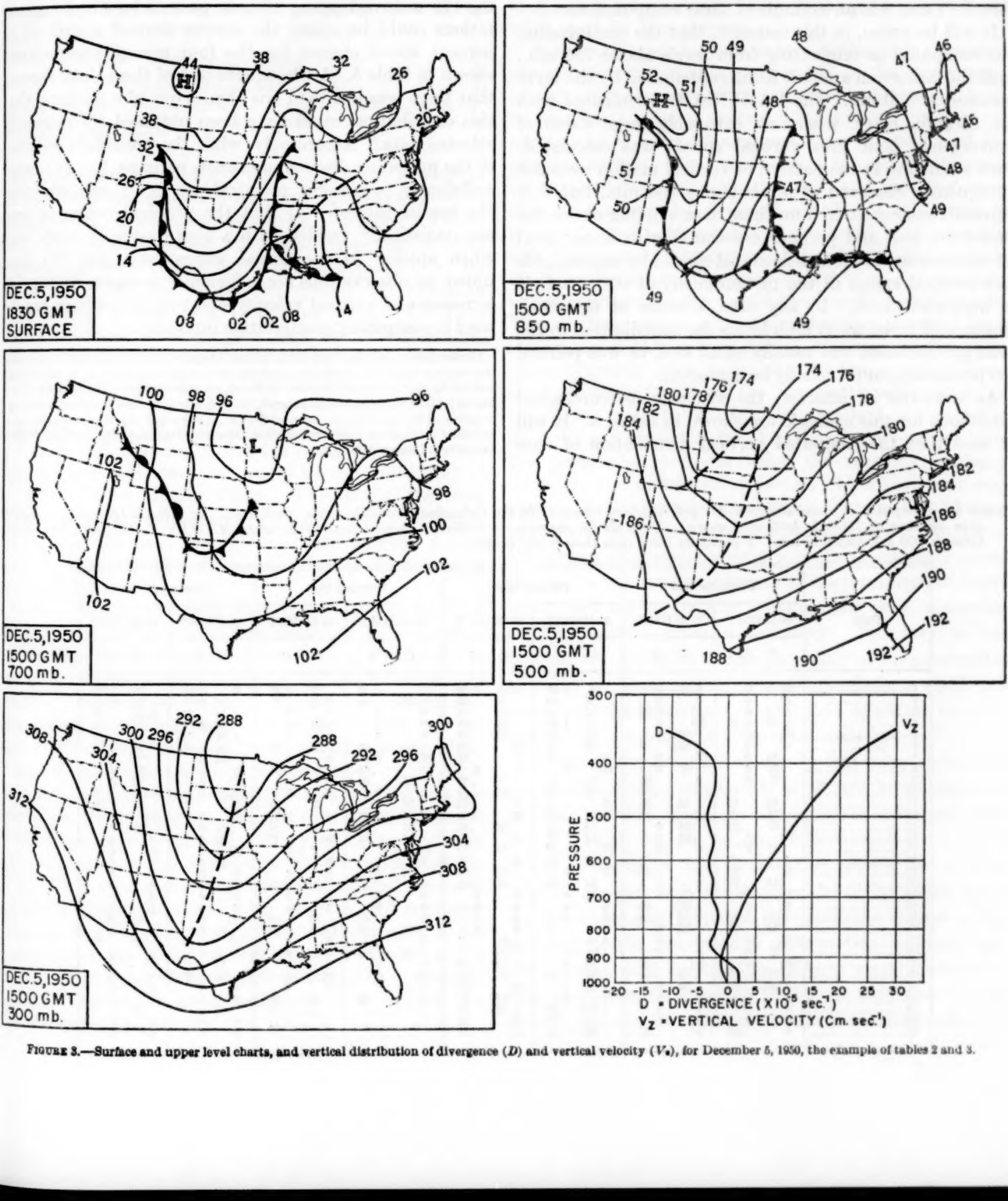

FIGURE 3.—Surface and upper level charts, and vertical distribution of divergence (D) and vertical velocity (V_z), for December 5, 1950, the example of tables 2 and 3.

dew point temperatures (>15° C.) and large vertical speeds (>5 cm. sec.$^{-1}$), errors as large as 15 percent or greater may be introduced. In such cases actual condensation temperatures and pressures were computed. Errors caused by these assumptions always cause the computed amounts to be higher than if actual condensation temperatures and pressures were used.

Using these assumptions, a procedure for computing the rate of rainfall for each layer may be programmed and the summation indicated by equation (2) performed in order to give the total precipitation during the 12-hour period. Table 3 is an example of these computations.

It will be noted, in this instance, that the precipitation was computed as originating from levels above 750 mb., with the maximum amount being contributed by the layer centered at 600 mb. The RAWIN data terminated with the 350-mb. layer, where an extremely large value of convergence (and thus vertical speed) was calculated. Such anomalies in the pattern of vertical motion were not infrequently observed at levels above 500 mb., but it is impossible to conclude from these data whether or not the effects are real and perhaps associated with upper level jet phenomena, or are fictitious and caused by unavoidable instrumental errors in the measurement of strong winds at high elevations. In any case, because of the small amounts of moisture at high levels the contribution to the total precipitation was usually small and, as was pointed out previously, could usually be neglected.

As a matter of interest, the synoptic meteorological conditions for this example are shown in figure 3. It will be seen that the computed vertical distribution of convergence-divergence and vertical motion associated with this situation seems quite reasonable when compared with the synoptic models usually thought to apply in the atmosphere.

SUMMARY OF RESULTS

Table 4 presents the results of carrying out these computations for the winter months of December 1949, and January, February, and December 1950. These data are summarized in table 5 for all cases where RAWIN-RAOB data were available above the 700-mb. level. Considering the 179 nonoverlapping 12-hour periods for which computations could be made, the results showed a skill of 27 percent above chance for the four precipitation classes shown in table 5. A Chi-square test of these data showed that there was less than one chance out of a hundred that this distribution could have been obtained by randomly selected data.[3] Accordingly, while the procedure outlined in the preceding discussion cannot, perhaps, be considered sufficiently precise for practical use in its present form, the results indicate (*a*) that the physical concepts used are statistically verified with a high degree of confidence when applied to the actual atmosphere, and (*b*) that upper air observations are sufficiently accurate to provide a measure of vertical velocity and moisture which can be used in computing precipitation intensity.

[3] It should be noted that the validity of the Chi-square test when applied to these data is somewhat doubtful because of the probable serial correlation between successive observations, as well as for other reasons. However, the auto-correlation coefficient for a one-period (12 hour) lag in the observed precipitation is only 0.4 so that this effect is not likely to influence the conclusion to any great degree. In any event, this serial correlation does not affect the magnitude of the relationship (correlation) between computed and observed precipitation shown in table 5.

TABLE 4.—*Computed (C) and observed (O) precipitation amounts in the Columbia-Nashville-Little Rock area. M=RAWIN and/or RAOB data missing at 700-mb. level and above. Amounts in parentheses indicate computations made when RAWIN/RAOB data terminated between 700 mb. and 500 mb. T indicates trace (less than 0.005 inch)*

Date	December 1949				January 1950				February 1950				December 1950			
	03-15 GMT		15-03 GMT		03-15 GMT		15-03 GMT		03-15 GMT		15-03 GMT		03-15 GMT		15-03 GMT	
	C	O	C	O	C	O	C	O	C	O	C	O	C	O	C	O
1	0	0	0.27	0	T	0.30	0.06	0.09	(0)	0.42	M	0.07	0	0.01	M	T
2	(0)	0	0	0	M	.63	M	.69	(0.07)	.05	0.01	T	M	0	M	0.41
3	0	0	(0)	0.03	M	.14	M	.65	0	0	0	0	M	.33	0.06	T
4	0.22	0.20	(0)	0	M	1.24	M	.44	(0)	0	(.02)	0	M	0	M	.02
5	(0)	0	M	0	M	.33	M	.03	0	0	0	0	0.08	.03	.24	.15
6	0	0	(.15)	.13	M	.21	(0)	.06	(.14)	.01	.01	.01	.14	.33	M	.06
7	M	0	(0)	0	0	.01	0	.01	(0)	0	0	0	M	0	(T)	0
8	.01	0	T	0	(0)	0	0	0	0	.01	.31	.04	(0)	0	T	0
9	T	0	0	.01	M	T	M	.02	M	.08	0	0	(0)	0	M	0
10	0	.14	0	.41	M	.89	M	.01	0	0	0	0	(0)	0	(0)	T
11	M	.94	M	.64	0	.01	0	.06	M	0	T	.04	0	T	.14	.03
12	M	.13	(0)	.19	0	.36	(0)	.29	M	1.45	(0)	1.21	(0)	0	(0)	0
13	.58	.02	.06	0	M	.57	M	.52	.09	.94	.16	.28	.04	0	(.06)	0
14	(0)	0	.06	0	M	T	0	T	M	.24	(0)	T	M	0	0	0
15	T	0	0	0	(0)	.26	M	.39	(0)	T	T	0	(0)	0	(0)	0
16	0	T	0	T	M	.02	0	T	0	0	0	0	M	0	0	0
17	0	.03	0	.56	0	0	M	.04	M	0	0	0	0	0	0	0
18	M	.24	0	0	M	.05	.08	.03	0	T	M	.01	(0)	0	0	0
19	0	0	T	0	M	0	0	0	0	0	0	0	0	0	0	0
20	0	0	(.02)	.01	0	0	0	0	0	0	0	0	0	0	0	0
21	(.05)	.04	M	.33	0	0	0	0	0	.07	.37	.39	(0)	0	.01	0
22	M	.27	M	0	0	.01	0	0	M	.27	(.05)	T	.08	0	0	0
23	(0)	0	0	0	0.06	T	.04	T	0	0	0	0	(0)	T	T	T
24	T	0	0	0	.05	0	M	T	M	.01	0	0	0	0	(0)	0
25	(0)	0	(0)	.02	M	0	M	.02	0	0	0	0	M	0	(0)	0
26	(.43)	.63	.30	.01	M	1.03	.25	.32	(0)	0	(0)	0	M	T	(0)	0
27	(.06)	.02	M	0	M	0	(0)	.01	0	0	0	0	0	0	0	0
28	0	0	0	0	0	0	M	T	0	.38	M	.16	0	0	0	0
29	0	0	0	0	(.02)	.04	(.30)	T	----		----		M	0	T	0
30	0	0	0	0	M	.05	(0)	.47	----		----		0	0	0	0
31	0	.02	0	.03	M	.16	(0)	.59	----		----		0	0	0	0

TABLE 5.—*Summary of computed vs. observed precipitation amounts from table 4. Data include all cases except those with missing (M) RAWIN observations.*

		Computed				
		0	0.01–0.25	0.26–0.50	>0.50	Total
Observed	0	110	14	2	0	126
	0.01–0.25	19	14	2	1	36
	0.26–0.50	8	3	1	0	12
	>0.50	3	1	1	0	5
	Total	140	32	6	1	179

Percent correct: $\frac{125}{179}=0.70$

Skill score: $\frac{125-105}{179-105}=0.27$

Chi-square (1 degree freedom): 20.8*

*This value was computed for a 2-by-2 table in order to combine frequencies so that expected values would exceed 5.

An examination of table 4 will show that, on many of the days with moderate or heavy precipitation, no computations could be made, due primarily to missing RAWIN data. By far the greatest number of these missing wind observations were caused by "limiting angles", i. e., by the balloon dropping below the critical 15° elevation angle where present RAWIN equipment (SCR–658) cannot be relied upon to give accurate measurements. From a study of individual ascents, it appeared that these limiting angles were due to (*a*) strong winds aloft predominantly from a single general direction and/or (*b*) slow ascensional rates, the latter probably being caused partly by lack of sufficient initial free lift being imposed on the balloon by the observer, and partly by the added drag of the precipitation during flight. This study therefore emphasizes both the importance of care in making observations and the need for improved instrumental equipment to alleviate the technical difficulty in obtaining winds aloft measurements.

It will also be observed that there appears to be a tendency for the computed amounts to be somewhat smaller than those observed. This bias is probably due in part to the neglect of such factors as radiational cooling of clouds, or to the fact that only large-scale precipitation mechanisms are considered. This latter effect is of particular interest, for it can be shown rather simply that, if upward small-scale motions are accompanied by compensatory downward motions within an area of large-scale circulation, the average precipitation over the area covered by the large-scale circulation will always be greater than that accounted for by considering the large-scale effects alone. This means that the computed rainfall amounts associated with thunderstorms, for example, will in general be too small, even when averaged over a large area, if the computations are based on the large-scale features. In an attempt to take this effect into account, the writers are currently engaged in an investigation of data from the Thunderstorm Project [13].

CONCLUSION

In the preceding discussion there has been described a method of attack on the problem of computing, from the currently observed fields of moisture and wind, the expected intensity of precipitation during the 12-hour period between RAWIN-RAOB observations. A physical, and therefore quite general, approach has been developed. A test of the procedure applied to the actual atmosphere has provided statistical evidence of the usefulness of the method, the validity of the assumptions made, and the accuracy and representativeness of the observational data when used for this purpose.

It should be pointed out, however, that this discussion represents only an exploratory study of the larger problem of computing precipitation from currently made prognostic charts. Such prognoses at present contain, in addition to the frontal analyses, only the height of the pressure surface. This means that the wind field must be obtained from an approximation, and that the moisture field cannot be obtained at all. It is desired to point out that a knowledge of the latter is of considerable importance in precipitation forecasting and a study of its prediction and eventual inclusion on the regularly issued prognostic charts should accordingly be undertaken. The effect of using a wind approximation obtained from the contour field is not yet known, but work on this problem is currently being undertaken by the writers. These studies include the use of the "gradient" wind instead of the "true" wind in the preceding method of computing divergence, a procedure for obtaining the divergence from the geostrophic vorticity using the vorticity tendency equation, and certain other devices which have been suggested. Preliminary results indicate that the first of these procedures is of little practical usefulness, but the use of the geostrophic vorticity in the vorticity tendency equation seems to provide estimates of the precipitation intensity which are, at least in a preliminary way, somewhat superior to those given in this paper. This may, as was pointed out earlier, be due to the ability of the geostrophic vorticity to filter out small scale circulations which are present in the "true" winds, and thus to reduce the random errors introduced by the small scale "noise."

Although the integration was, in this case, carried out for 10 layers using increments of 50 mb. in thickness, it is probable that useful results could be obtained from fewer layers. In an incidental investigation, using a group of the same data included in table 4 but making calculations with only the standard levels for which upper air charts are now prepared, no significant change in the accuracy of the precipitation computation was observed. This means, that, from an operational standpoint, accurate prognostic information limited to, say, 4 or 5 levels would probably be sufficient to provide useful results.

ACKNOWLEDGMENT

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ESTIMATING PRECIPITATION AT SAN FRANCISCO FROM CONCURRENT METEOROLOGICAL VARIABLES[1]

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ABSTRACT

Variables obtained from synoptic sea level and upper air charts are investigated to determine their significance in the estimation of concurrent rainfall. Eight variables consisting of sea level pressures and pressure gradients, pressure heights and height differences, and the temperature-dew point differences at two upper levels are combined into a graphical procedure to estimate the probability of occurrence of rainfall. With the probability of occurrence rising to above 50 percent, supplementary charts are used to estimate the amount of rainfall to be expected.

INTRODUCTION

In making a forecast the usual procedure is to predict the movement and expected change in intensity of pressure systems, with their associated fronts, and the expected positions and intensities of upper air troughs and ridges [1]. Although these expected changes are often treated in a subjective manner, the final stage of such a forecasting procedure may well lead to the actual construction of prognostic charts of the synoptic conditions expected at some period in the future. However, experience has indicated that even if accurate prognostic pressure and upper level contour charts were available, forecasters would still not be able to give a complete picture of the accompanying weather conditions. Thus, a study leading to an increase in the knowledge of the current weather to be expected from a given synoptic situation is of interest to the forecaster. In a study of this type, the synoptic charts take the place of the prognostic charts which are expected to be used eventually in the forecasting system. Such a study will furnish information as to the significance of commonly observed variables and indicate the accuracy which may be attained through their use.

Of additional interest is the possibility that a study of this nature may lend itself to the evaluation of attempts to modify the weather. During the past few years the problem of evaluating the effect of cloud seeding on the production of rainfall has become of considerable importance. This problem has no easy solution. The difficulty lies largely in the fact that the areal distribution of rainfall for the individual storm has great variability, and further that the final solution lies in the realm of statistics where a definite yes or no answer is not forthcoming. It is fairly obvious that if a method could be devised which would establish a perfect correlation between meteorological variables and the associated rainfall at a given point, the problem of evaluating rainmaking at that particular point would be solved—on the condition that the seeding itself had no influence on the meteorological variables (a condition perhaps not fulfilled for all variables used in this study). The nature of the weather problem precludes the attainment of perfect correlation and the eventual value of the derived method would depend upon the degree of correlation which is obtained.

GENERAL ASPECTS OF THE RAIN PROCESS

Meteorologists are not in full accord as to the prerequisite conditions for the occurrence of rain. Although knowledge of the precipitation mechanism is increasing, a great deal is yet to be learned about the exact physical conditions leading to rainfall. In the consideration of factors favorable for the production of rainfall, the most common agreement appears to be that upward motion of the air and the presence of sufficient moisture are of predominant importance [2]. Of less general agreement is the extent to which the occurrence of rain depends upon the temperature distribution within the cloud system, the number and nature of nucleating particles, the drop-size distribution of the cloud, and various other factors.

In estimating the occurrence and amount of precipitation the meteorologist is faced with the fact that, with the possible exception of the moisture, the physical factors entering into the production of rain are not measured directly. The existence of large-scale vertical motions in the atmosphere must be inferred from the analysis of the major air currents [3]. No information is available as to the number of particles in a given airmass which may serve as nucleating or sublimating agents, nor is there any synoptic information concerning the distribution of drop-

[1] Paper presented at 116th National Meeting of the American Meteorological Society, Corvallis, Oreg., June 16-18, 1952.

size within a particular storm cloud. Accordingly, an estimate of the effects which the existing combination of physical factors has on the production of rainfall must be obtained indirectly. In practice, it is the usual procedure in making a forecast to go directly from the prognosticated synoptic features to the expected weather, thus bypassing the actual consideration of the quantitative effects of the physical factors. However, a thorough understanding of the physical process may suggest significant meteorological variables for use in a forecasting system. A considerable number of variables suggested by theory and synoptic experience were investigated. In general, it was found that variables involving pressures and pressure heights were more effective in determining rainfall at the station studied than were those involving temperature and stability. Only those variables showing the greatest effectiveness were used in the estimating procedure which was developed.

PURPOSE OF STUDY

This study was carried out in order to investigate the effectiveness of commonly observed synoptic factors as indicators of concurrent rainfall at San Francisco and to gain a better understanding of the part which these factors play in bringing about a rain-producing situation. The final stage of the study was the combination into a graphical procedure of the most effective factors in order to obtain an estimate of the rain-producing potentialities of a given synoptic situation. The availability of the necessary prognostic charts would lead to an objectively determined forecast based on the prognostic charts.

PROCEDURE

In order to investigate the factors responsible for rain at San Francisco, combinations of meteorological variables were studied by means of scatter diagrams on which were entered the occurrence or nonoccurrence of rain during a period centered at about the time of the meteorological observations. Data from the 1030 and 2230 PST sea level charts and 0700 and 1900 PST radiosonde and pilot balloon flights were used in connection with the rainfall occurring between 0430 and 1630 PST for the morning data and between 1630 and 0430 PST for the evening data. Thus, for this portion of the study, the sea level variables were taken at the midpoint of the 12-hour precipitation period and the upper air data about 3 hours after the beginning of the period. In as much as traces of precipitation may occur from either a rain situation or a fair weather fog or stratus condition, all traces were omitted from consideration in this portion of the study. In developing the quantitative aspects of the study, sea level data taken every 6 hours were correlated by means of a scatter diagram with the precipitation occurring during the 3 hours prior to and the 3 hours immediately following the observation. In this part of the study, traces were counted as no rain. The months of November through March of the 1950–51 and 1951–52 seasons were used in the development of the procedure and the 1948–49 and 1949–50 seasons were used as a test.

In analysing the scatter diagrams involving the meteorological variables, the smoothness and uniform spacing of the lines of equal probability were considered of primary importance. The general shapes and positions of the lines were determined by means of the methods of analysis presented by Brier [4] and Kangieser and Jorgensen [5]. The final positions were checked by determining the ratios of the rain to no-rain cases between the lines. However, if the attainment of the correct ratios was impossible without distorting the lines, the smoothness of the lines was allowed to take precedence over the correct ratio. This smoothing has the effect of assuming that large random variations may occur in the ratios in some portions of the chart due to the small amount of data used in determining these ratios. In analysing the scatter diagrams involving two probabilities, the formula given in [5] was used to describe the shapes of the lines (e. g., see fig. 5) with the positions of the lines adjusted to bring the channel frequencies more into harmony with the expected mean values. However, here again the smoothness of the lines and the symmetry of the configuration were not sacrificed. In general, the end charts were not believed to have sufficient data in the central portions to place the lines in this area with a high degree of accuracy. In analysing the quantitative charts (figs. 8 and 9), the general shapes of the lines were determined by inspection. Once the shapes of 2 or 3 random lines throughout the data were determined, lines of specific values were then located in such positions to give approximately the correct mean values of the plotted amounts in the areas between the lines. Here again, smoothness of the lines and uniform spacing were maintained at the expense of the channel means.

In order for the test of the procedure to be impartial and independent of the original development, the test was not carried out until the study was completed. Final probabilities and estimated amounts were obtained for all the test data and entered on the tabulation sheet before the actual amounts were entered.

CHOICE OF METEOROLOGICAL VARIABLES

Heavy rain in central California usually results from the occurrence off the coast of a cyclonic development in connection with associated occluded frontal systems, while lighter amounts of briefer duration are common as the result of frontal passages moving in from the west or northwest. In the case of heavy rains, the causative storm may vary greatly in size and mode of origin, with the movement into the area off the coast from almost any direction, but most generally from the southwest or west. The larger storm systems with their accompanying troughs may bring moist air into the area from tropical or subtropical latitudes and if persistent may lead to flood conditions. Storms moving into the coastal area from a northerly di-

rection may have insufficient moisture to produce rain if the path is over land, but if over water, sufficient moisture may be accumulated in the storm to cause heavy rain. Cyclonic development along a frontal system approaching the coast may result in locally heavy rain, with heavy rain at San Francisco occurring when the center of the lowest pressure approaches the coast to the north of the station with the lowest pressure at Fort Bragg, Calif., or Eureka, or occasionally at San Francisco. Once the center of the Low has reached a position to the east or south of the station, rain usually ceases rather abruptly. Occasionally weak disturbances undergo rapid deepening as they approach the coast, the deepening apparently the result of conditions becoming favorable for storm development throughout the entire troposphere. Light rain of brief duration may occur under various other conditions.

A unique feature of the rainfall in California is the readily observable effect of the topography on the distribution and amount of rain from a given storm. The coastal mountain range and the Sierra Nevada act as permanent upglide surfaces. Thus, westerly rain-bearing winds are forced to rise with the resulting lifting becoming a predominant factor in the production of rainfall. As a consequence, the strength of the westerly flow over the area and the moisture content of the air become significant variables.

The search for meteorological variables was made on the assumption that vertical motion due to convergence and orographic lifting together with a sufficiently high moisture content of the air were adequate to account for the observed frequency and quantity of rainfall. Variables from the sea level and upper level charts were investigated. Widespread upward motions in the lower troposphere were assumed to be associated with the circulation about low pressure systems and widespread downward motions with the circulation about high pressure systems. Accordingly, the relationships between various sea level pressures and the corresponding weather were studied. Observation of the position of the jet stream during the winter season indicates that significant precipitation usually does not occur unless a zone of maximum westerlies at 500 mb. (lower portion of the higher jet stream) has migrated southward to a position near the station [6].

Regardless of the manner in which the rain situation develops, this study has shown that certain rather definite meteorological conditions need to be fulfilled in order for significant rain to occur. Among these conditions are: (1) Low pressure along the California coast at or to the north of San Francisco and relatively higher pressure to the south, a condition resulting from the presence of a storm offshore and leading to strong westerly flow over the coastal and interior mountains, (2) Sufficiently high moisture content of the air from the surface to about 10,000 feet, indicating that the previous history of the airmass was such as to allow the air to pick up a supply of moisture, and (3) Circulation in the upper air such as to give a zone of maximum westerly winds just to the north of the station with relatively low pressure-height values to the north and high values to the south.

PROCEDURE FOR ESTIMATING PROBABILITY OF RAIN OCCURRENCE

In order for the factors listed above to be taken into account in the estimation of rain occurrence, the following pairs of variables were incorporated into the estimating procedure:

(*a*) The sea level pressure difference, Santa Maria minus Fort Bragg, Calif., against the sea level pressure at San Francisco as shown in figure 1. The pressure difference furnishes evidence of the onshore flow, and the sea level pressure at San Francisco indicates the nearness or intensity of a low pressure system.

(*b*) The temperature-dew point difference at 700 mb. at Oakland against the same variable at 850 mb. These variables are plotted in figure 2. A small temperature-dew point difference indicates the nearness to saturation of the air or perhaps the actual existence of clouds. Once clouds have formed through a rather deep layer, a small amount of convergence or lifting will greatly increase the likelihood of rain.

(*c*) The height difference in the 500-mb. surface between Oakland and Medford against the difference between Santa Maria and Oakland as given in figure 3. Observation during several winter seasons indicates that a significant aspect of the upper air charts in determining the likelihood of rain at San Francisco is the position and strength of the zone of maximum winds above 10,000 feet. As shown in figure 3, a majority of rain cases occur with the height difference at 500 mb. between Oakland and Medford of 200 feet or greater with a somewhat lower value between Santa Maria and Oakland. This combination of variables brings into the estimating procedure the effect of the upper air flow.

(*d*) The height of the 500-mb. level at Medford against the sea level pressure at Eureka as shown plotted in figure 4. With the movement of storm centers onto the Washington and Oregon or extreme northern California coast, or the approach of fronts from the northwest, the sea level pressure at Eureka becomes a significant variable for the estimation of the occurrence of rain at San Francisco. A previous study [7] has shown that the significance of low pressure at Eureka is dependent on the circulation aloft. The lower pressures at Eureka are not so strongly indicative of rain at San Francisco when the 500-mb. height at Medford is relatively high. Thus, the combination of the 500-mb. height at Medford with the sea level pressure at Eureka increases the effectiveness of this latter variable.

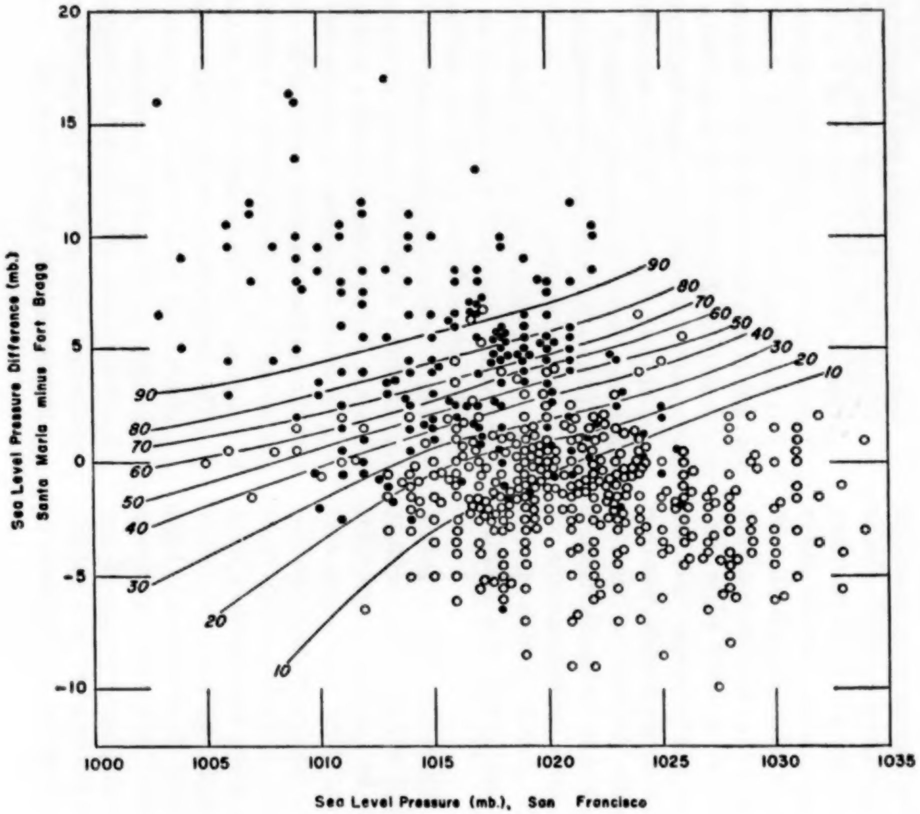

FIGURE 1.—Scatter diagram combining the sea level pressure at San Francisco and the difference in sea level pressure between Santa Maria and Fort Bragg, Calif. Circles represent no-rain cases and dots rain cases at San Francisco Airport with the observational period covering a 12-hour interval from 0430 to 1630 PST and from 1630 to 0430 PST. The pressure data are taken from the 1030 and 2230 PST sea level observations. Probability P_1 is read from this chart.

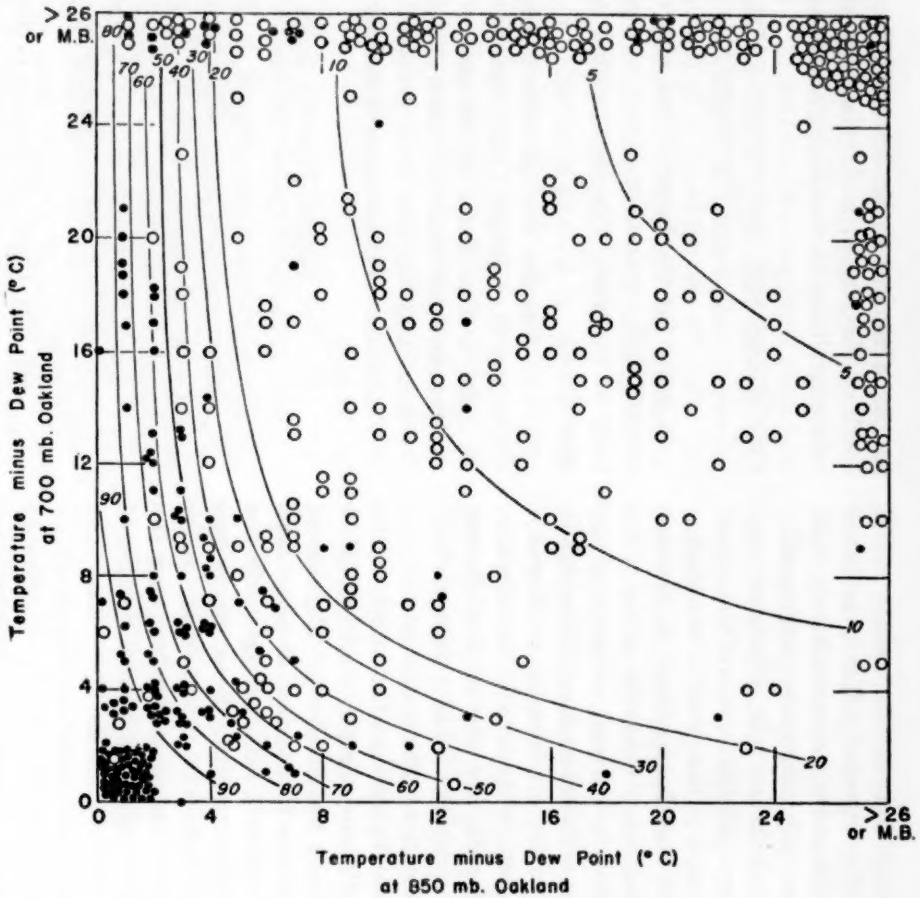

FIGURE 2.—Scatter diagram combining the temperature-dew point difference at the 850- and 700-mb. levels at Oakland in terms of the occurrence of rain or no-rain as in figure 1. The temperature and dew-point data are taken from the 0700 and 1900 PST radiosonde observations. Probability P_2 is read from this chart.

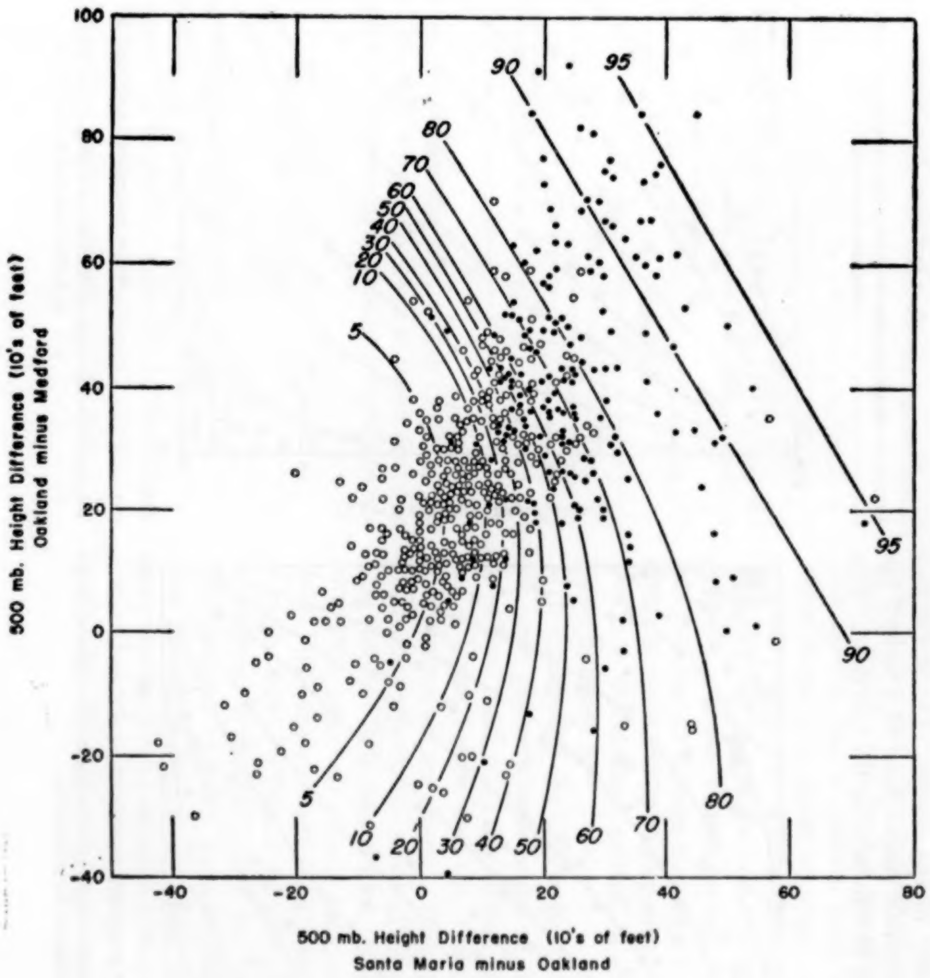

Figure 3.—Scatter diagram combining the height difference in the 500-mb. surface between Santa Maria and Oakland and between Oakland and Medford in terms of the occurrence of rain or no-rain as given in figure 1. The 500-mb. data are taken from the upper air observations as in figure 2. Probability P_3 is read from this chart.

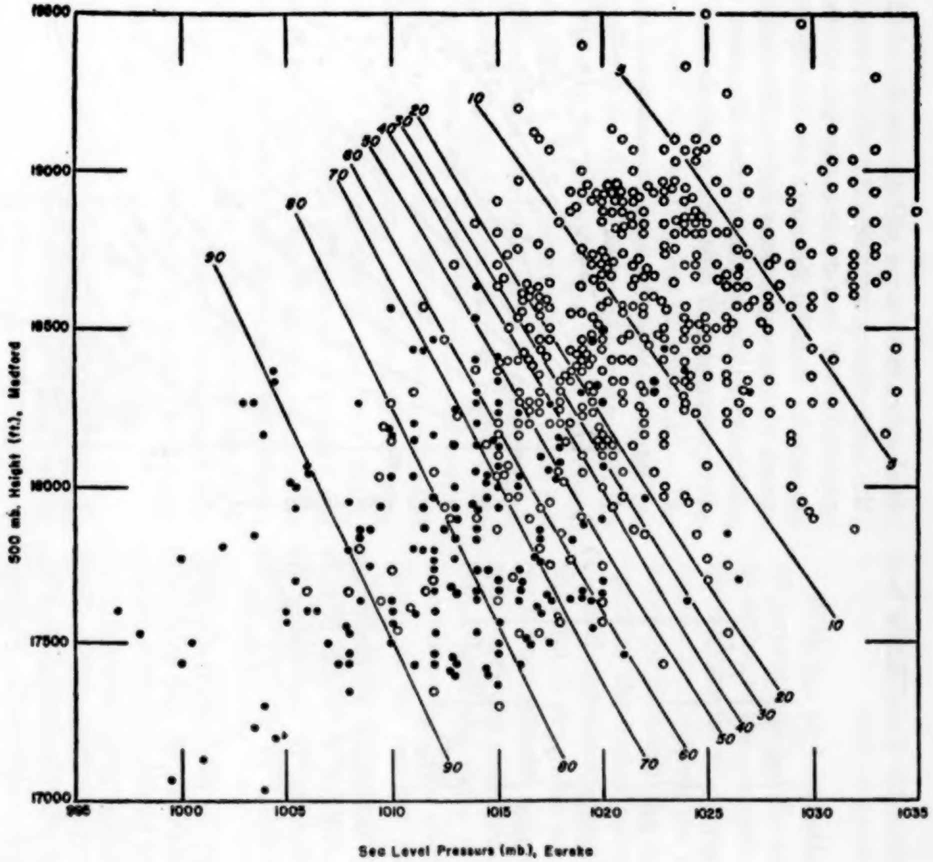

Figure 4.—Scatter diagram combining the sea level pressure at Eureka and the height of the 500-mb. level at Medford in terms of the occurrence of rain or no-rain as in figure 1. The data are taken from the observations as indicated in figures 1 and 2. Probability P_4 is read from this chart.

The four scatter diagrams thus developed have been combined according to the following outline:

Sea level pressure difference, Santa Maria minus Fort Bragg.	P_1 (fig. 1)	$P_{1,2}$ (fig. 5)	W (fig. 7)
Sea level pressure, San Francisco			
Temperature minus dew point at 700 mb., Oakland.	P_2 (fig. 2)		
Temperature minus dew point at 850 mb., Oakland.			
500-mb. height difference, Oakland minus Medford.	P_3 (fig. 3)	$P_{3,4}$ (fig. 6)	
500-mb. height difference, Santa Maria minus Oakland.			
500-mb. height, Medford	P_4 (fig. 4)		
Sea level pressure, Eureka			

In the scatter diagrams the combined variables are evaluated in terms of the probabilities, P, which in turn are combined into the final parameter, W, expressing the overall probability for rain to occur as a result of the existing variables.

The estimate of the probability of rain occurrence may vary over the full range of values from near zero to close to 100 percent. For the purpose of expressing the accuracy of the various charts the percentage of correct estimates based on the 50 percent line is used. Counting as errors the no-rain cases falling above the 50 percent line and the rain cases falling below the line, percentages are obtained representing the accuracy of the charts. For the initial charts, the percentage correct ranges from 86.5 to 88.3 for the dependent data and 79.6 to 90.6 for the test data. The final chart gives a percentage correct of 92.9 and 92.8 for the respective groups of data. In general, the higher the probability of occurrence the greater is the expected amount. Rainfall amounts are usually light when they occur with an expectancy below 50 percent.

An inspection of the scatter diagrams leading to the final parameter W shows a gradual accumulation of the

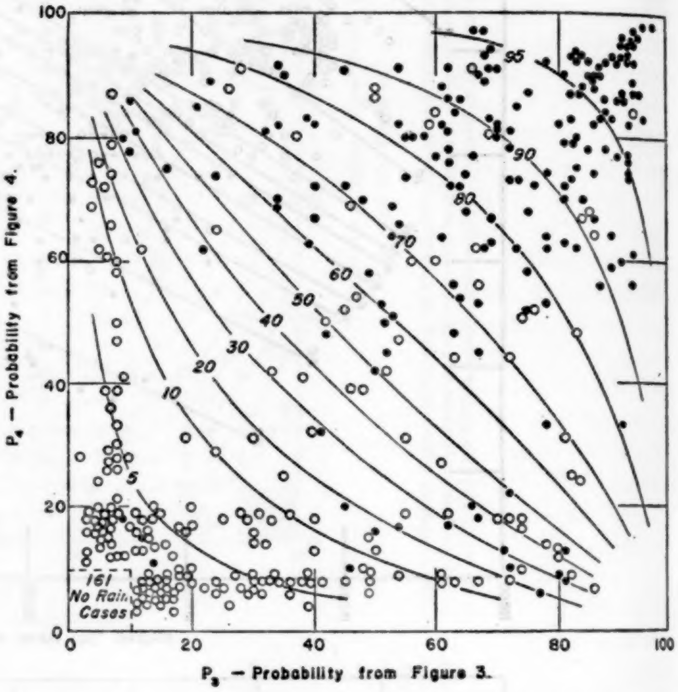

FIGURE 6.—Scatter diagram combining the probabilities obtained from figures 3 and 4 and giving the probability $P_{3,4}$.

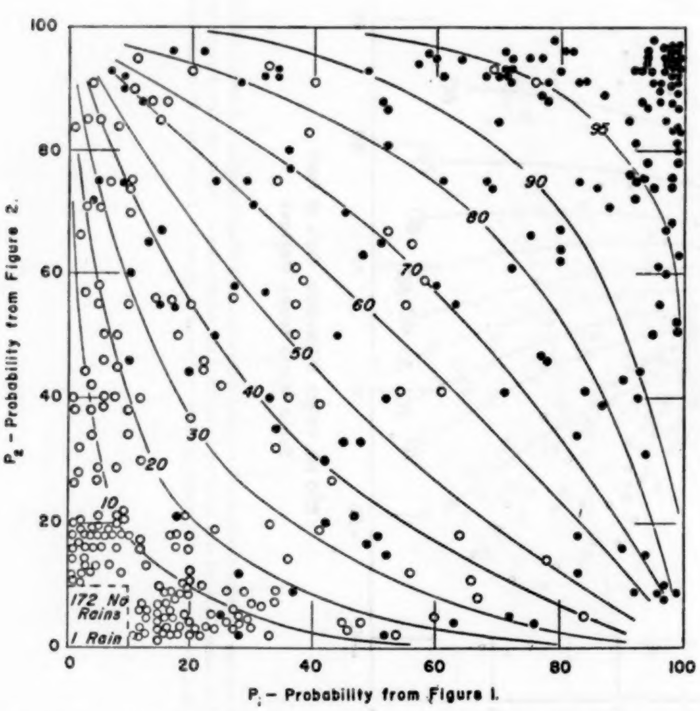

FIGURE 5.—Scatter diagram combining the probabilities obtained from figures 1 and 2 and giving the probability $P_{1,2}$.

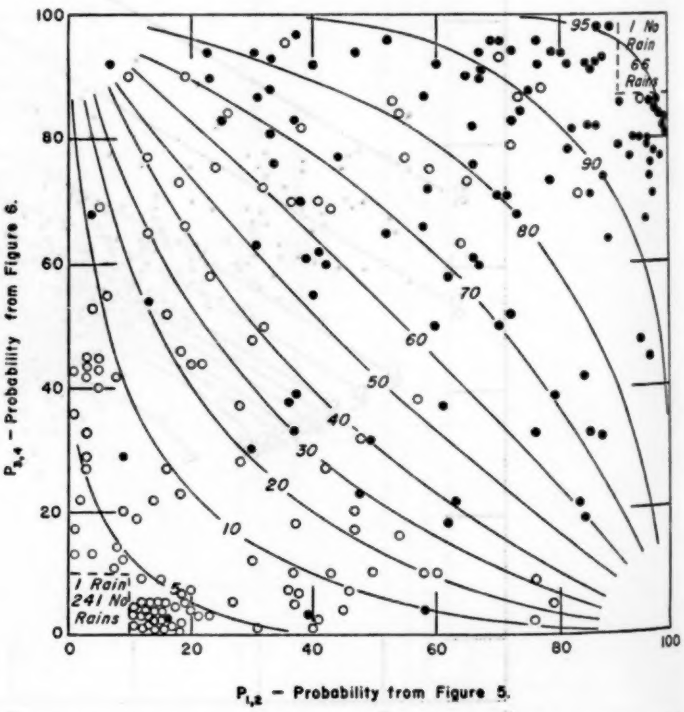

FIGURE 7.—Scatter diagram combining the probabilities obtained from figures 5 and 6 and giving the final probability W.

rain cases in the high probability areas of the charts and the no-rain cases in the low probability areas. This shifting of the cases to the upper right- and lower left-hand corners suggests that in the perfect final chart all the rain cases would fall at 100 percent probability and all the no-rain cases at zero percent probability. (At this point there would be no useful relationship between the probability of occurrence and the expected amount.) The degree to which the contrasting weather types are shifted to opposite corners of the charts becomes a measure of the confidence which may be placed in the use of the chart. For example, in figure 1, 40.2 percent of the rain cases fall above the 90 percent line compared to 59.2 percent which fall above the same line on the final chart in figure 7, the gain of 19 percent resulting from the additional sets of variables. Similarly, 68.6 percent of the no-rain cases fall below the 10 percent line in figure 1 compared to 84.9 percent which fall below the same line in the final chart. Listed in table 1 for the charts in figures 1 through 7 are the percentage of correct cases based on the 50 percent line, the percentage of rain cases above the 90 percent line, and no-rain cases below the 10 percent line for both the dependent and test data. As may be seen, there is a gradual improvement in both features up to the final chart.

TABLE 1.—*Tabulation of the percentage of estimates correct based on the 50 percent line, the percentage of rain cases above the 90 percent line, and the no-rain cases below the 10 percent line. (Data based on 12-hourly periods.)*

A. Dependent data (537 cases)

	Scatter diagram						
	P_1	P_2	P_3	P_4	$P_{1,2}$	$P_{3,4}$	W
Percentage correct	88.3	86.6	86.3	88.1	91.8	90.0	92.9
Percentage of rain cases above 90% line	40.2	34.5	17.7	21.5	50.6	51.9	59.2
Percentage of no-rain cases below 10% line	68.6	64.1	59.2	61.6	75.3	79.4	84.9

B. Test data (501 cases)

	Scatter diagram						
	P_1	P_2	P_3	P_4	$P_{1,2}$	$P_{3,4}$	W
Percentage correct	87.6	90.6	79.6	85.4	92.0	86.6	92.8
Percentage of rain cases above 90% line	27.7	42.4	18.2	13.6	47.0	44.7	56.1
Percentage of no-rain cases below 10% line	79.3	65 3	53.6	61.5	85.4	73.6	86.7

The skill score, based on the 50 percent line for the final chart, W, for the dependent data is 0.843 and for the test data 0.812. The score is obtained from the expression

$$\text{Skill score} = \frac{C - E_c}{T - E_c}$$

in which T=number of estimates made

C=number of estimates correct

E_c=number of estimates expected to be correct with the estimates distributed at random over the period covered by the data.

PROCEDURE FOR ESTIMATING PRECIPITATION AMOUNTS

With the probability of rain occurrence rising to above 50 percent (the usually assumed value), it is then necessary to estimate the amount of rain to be expected from the given synoptic situation. Although there is a significant correlation between the probability of occurrence and the expected amount, the relationship has decreased significance at the higher values of the probabilities. This decreased significance results from the fact that once the situation is favorable for "heavy" rain, the probability of occurrence remains nearly the same even though the heavy rain may vary considerably in amount. In addition, it is found that the value of W may vary greatly during a 12-hour period, the change sometimes amounting to as much as 90 percent between two successive periods. Under these changeable conditions observations every 12 hours are too infrequent to be satisfactory for use in the estimation of rainfall amounts. Variables which lend themselves to more frequent evaluations are desirable.

In evaluating the primary charts in terms of the estimation of rainfall amounts, the combination of variables involving the sea level pressure difference between Santa Maria and Fort Bragg and the sea level pressure at San Francisco were found to give the best quantitative results. Since these variables may be evaluated every 6 hours, they have been used in this part of the investigation. A study of the amounts expected reveals that when closed circulation prevails at 10,000 feet in the vicinity of the station, the surface variables chosen have somewhat different significance from a quantitative standpoint. For this reason, the situations have been divided into two types depending upon whether or not a closed Low is present at 700 mb. within the area bounded by the 115° and 130° W. meridians and the 25° and 45° N. circles of latitude. Those situations not involving a closed Low and making up a large majority of the cases are shown in figure 8, while the 10 to 15 percent of the situations during which a closed Low existed in the designated area are given in figure 9.

In developing the charts given in figures 8 and 9, the sea level pressure data taken at the 6-hourly map times have been used, with the rainfall amount occurring within the 3 hours before and the 3 hours after the observation plotted on the charts. Only those 6-hourly data are plotted for which the value of W is 50 percent or higher. The value of W obtained from the 1000 PST sea level and 0700 PST upper air data is assigned to the 1000 PST and 1600 PST observations, and similarly, the W obtained from the 2200 PST sea level and 1900 PST upper air data is assigned to the 2200 PST and the following 0400 PST observations. In this part of the study, when missing data prevented the full evaluation of W, the data available were used to give an estimate of the value of the parameter. Estimated 6-hourly amounts have been obtained for the test period, November through March 1948–49 and 1949–

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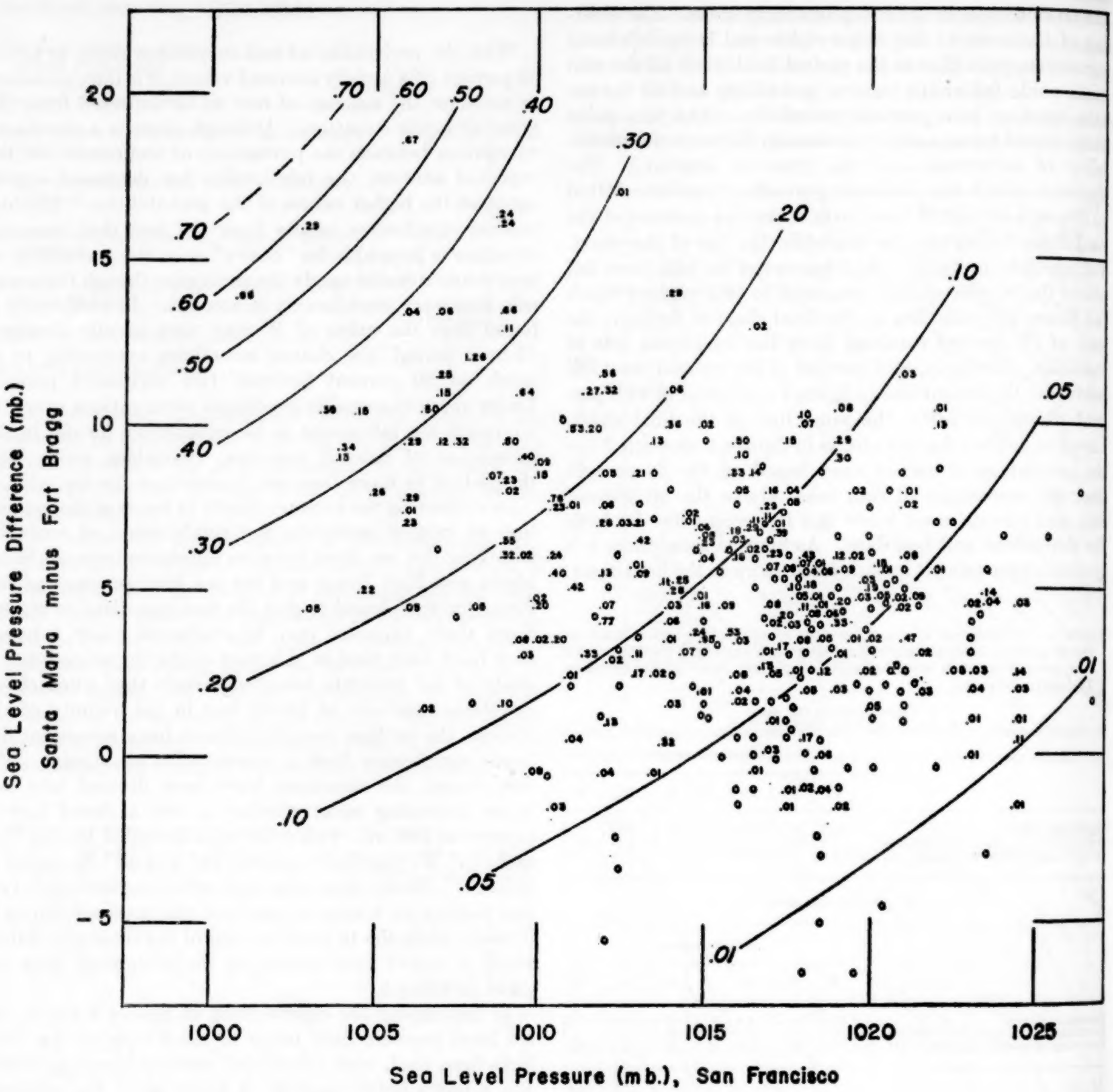

FIGURE 8.—Scatter diagram used for the estimation of 6-hourly precipitation amounts for those cases for which W is greater than 50 percent and no closed Low exists at 700 mb. in the designated area. (For designated area see text.) The data are taken from the 6-hourly observations with the rain period covering the interval within plus or minus three hours of the time of observation.

50. These amounts have been combined into 24-hour totals (24 hours beginning at 0100 PST) and are compared with the observed amounts in table 2.

CONCLUSION

For the particular station under consideration, the precipitation to be expected from a given synoptic situation may be estimated with worthwhile accuracy by means of sea level pressures, 500-mb. heights, plus a variable indicating the moisture content of the air. It is concluded that the procedure thus developed may be used as an objective aid in the interpretation of prognostic charts in terms of the expected weather. The substituiton of predicted values of the variables in place of the current values

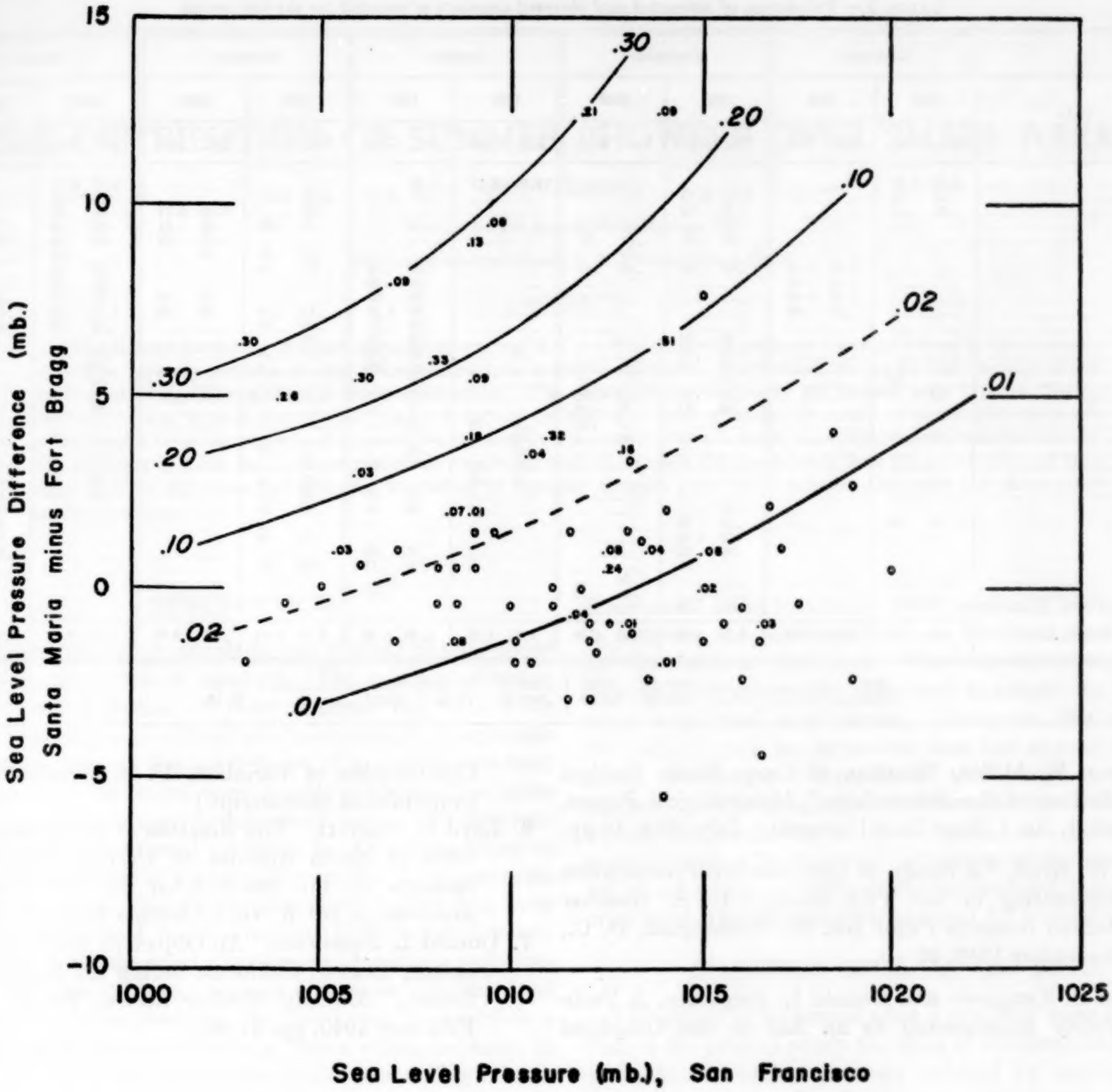

FIGURE 9.—Scatter diagram similar to that given in figure 8 except that a closed Low exists in the designated area.

will indicate the weather expected to accompany the prognosticated conditions. The success of this type of forecasting procedure lies in the accuracy with which the necessary variables may be predicted. The use of such a system for a trial period will determine its effectiveness.

Considerable variation is noted in the day-to-day agreement between the estimated and observed amounts given in table 2. This variation is unfavorable for the use of the procedure in evaluating the effects of weather modification efforts. However, the monthly and seasonal totals are sufficiently in agreement to promise some usefulness when the weather modification attempts are extended over a prolonged period.

ACKNOWLEDGMENTS

Appreciation is expressed to Mary E. Stoneback for her aid in the tabulation of the data and in drafting the final charts.

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TABLE 2.—*Tabulation of estimated and observed amounts of rainfall for the test period*

Date	November				December				January				February				March			
	1948		1949		1948		1949		1949		1950		1949		1950		1949		1950	
	Est.	Obs.	Est.	Obs.	Est.	Obs.	Est.	Obs.	Est.	Obs.	Est.	Obs.	Est.	Obs.	Est.	Obs.	Est.	Obs.	Est.	Obs.
1	0.20	0.03							0.09	0.12	0.03	----					0.37	0.13		
2	.14	.04			0.11	0.10					.06	0.15	0.13	0.27			.89	.62		
3	.05	.11			.10	.22							.20	----	0.39	0.72	.36	.17		
4					.01	----	0.04	0.11			.06	----	.28	.60	.61	.81	.06	.09		
5					.23	.21	.01	.07							.61	.66	.03	.09	0.05	0.05
6							.10	----					.30	.48	.14	----	----	.03		
7			0.09	0.20	.03	.05	.40	.88			.56	.34	.07	.02			.02	.01		
8			.41	.01							.51	.50					.01	.03		
9			1.31	1.29							.24	.05			.12	.16	.91	.43	.06	----
10			.38	.01							.62	1.23	.14	.18	.24	.26	1.18	.62	.01	----
11											.24	.24	.29	.31			.84	.70	.02	.01
12					.20	.09					.04	.19	.05	----						
13					.51	.23			.29	----	.42	.73					.12	----		
14					.32	.18	.02	.26			.17	.23					.14	----	----	.01
15	----	.01			.08	----	----	.11			.03	----					.56	.12		
16	.07	----			.60	.54	.12	.02			.52	1.04					.32	.01		
17					.06	.04	.54	.30			.21	.50					.08	----	.01	.04
18							.40	.61	.05	.05	.10	----					.31	.12	.06	.01
19	.04	.06					.04	----	.39	.67							.18	.67	.06	.10
20									.03	----	.02	----	.17	.02						
21									.28	.13	.02	.06	.02	----					.04	.02
22									.20	.27			.04	.01			.40	.41	.05	.14
23											.12	.03	.13	.17			.05	.03	.30	.86
24					.41	----					.10	.04	.18	.45				----	.47	.38
25	.02	.02			.24	.73													.15	.01
26					.28	1.44							.03	.09					.12	----
27					.06	.22					.35	.83							.01	----
28											.06	.11	.02	.01						
29							.02	----												
30							.01	----			----	.01								
31					----	.02														
Totals	.52	.27	2.19	1.51	3.24	4.07	1.70	2.36	1.33	1.24	4.48	6.28	2.05	2.64	2.11	2.61	6.83	4.28	1.41	1.63

Est.	1948–49	13.97	1949–50	11.89	Total	25.86
Obs.	1948–49	12.50	1949–50	14.39	Total	26.89

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THE DISTRIBUTION OF SUMMER SHOWERS OVER SMALL AREAS

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ABSTRACT

The spatial scattering of summer showers over much of the United States presents many problems to the forecasters. The primary purpose of this investigation is to determine the areal distribution, during 24-hour periods, of the occurrence of summer rains over three small areas. The areas selected for study are located near Lincoln, Nebr., Peoria, Ill., and in east central Ohio and are defined by circles of radii of about 35 miles. Precipitation data are taken mostly from the records of cooperative stations over a period of 4 to 6 years, depending upon available records. The analyses of these data are compared with a similar analysis of data for the Atlanta area [1] and it is found that there is little difference between the scattering of summer showers over the Southeast and over the three more northerly areas.

INTRODUCTION

Summer rains over much of the United States occur mainly in the form of showers. The average summer rainfall over a period of years changes but little over short distances, where there are unimportant topographic differences, but, on the other hand, it is well known that amounts or occurrences vary greatly within a few miles during periods or days of summer showers. For this reason, the term "scattered" is often used by forecasters to describe the expected shower distribution during specific periods or days. Shower forecast terminology has been receiving more and more attention during recent years, but research on the areal distribution of summer rain during particular days or periods has been limited. An attempt has been made in the Southeast to establish a forecast terminology containing terms which indicate, in effect, the forecaster's estimate of the probability of rain in any particular spot, such as on any given farm. It has been assumed that if the distribution of showers on any given day is more or less random, the probability of a shower at a given location will be indicated by the number of points within an area of reasonable size that actually get rain. This may or may not be a valid assumption. Furthermore, the exact size of an area of "reasonable size" and the optimum spacing of observing stations are open to question. It is beyond the scope of this paper to discuss either of these points. Nevertheless, it would be desirable to establish a shower terminology that would apply to the whole country, or at least to the area between the Rocky Mountains and the Appalachians. For this purpose it would be desirable to make studies similar to [1] covering, in sufficient detail, the whole country, or at least the area just mentioned. It is the purpose of the present work to make, at the suggestion of the Atlanta Research Forecaster, such a study for three small areas located in more northerly latitudes and to compare the results with those obtained by Beebe [1] for the Atlanta, Ga., area. The amount of time available makes it impossible at present to extend the study to cover additional small areas. Moreover, the necessity for making such an extension does not appear to be as great, considering the results of this study, as would be the case if large geographical differences in shower distribution had been found. It is hoped that the data presented herein will give some indication of the practicability of adopting a standard shower-forecast terminology, or at least provide some background for further study.

SELECTION OF AREAS AND DATA

This study is patterned after a previous work [1] except that in the present study the sizes of the areas and number of stations in each area were reduced by one-half. Examination of the distribution of precipitation-reporting stations [2] over several States along the 40th parallel and between the Rocky Mountains and the Appalachians suggested three small areas for study. These are located near Lincoln, Nebr., Peoria, Ill., and in east central Ohio, and are defined by circles of radii of about 35 miles. In determining the spatial distribution of summer rains, it would be desirable to use only rain periods or assign each shower or rain at each station to a particular date. However, data and time limitations preclude such a refined determination of shower distribution. A rain day here is defined as a date during which measurable rain is recorded during a 24-hour period. It would be desirable in this study to end the 24-hour period near the time of the minimum frequency of occurrence in order to minimize the number of apparent rain days due to the occurrence of rain from 1 shower during 2 consecutive 24-hour periods. At Atlanta, the frequency distribution of hourly occur-

rences shows a decided minimum about mid-morning and stations were selected for use in that study which measured 24-hour precipitation amounts around 0700 EST. It was not possible to select the reporting stations for the present study entirely on such a basis because not enough stations in these areas made observations at the proper times. Instead, the selections had to be based upon the greatest number of stations, within a particular area, that reported at about the same time. It was found that only 20 stations in each area could be utilized and even with this limited number it was necessary in some cases to use data from a few stations at which the observation time differed by about 6 hours. The lengths of record used for the study varied between 4 summer seasons for the Peoria area (1948–51) and 6 seasons for the Lincoln area (1946–51). More data were not readily available. There were occasional periods during which the record from a station was missing and data from some nearby station not included among the 20 were substituted.

RESULTS FOR THE PEORIA AREA

The 20 stations that were selected in the Peoria area reported 24-hour precipitation amounts ending near 1800 local time. The information readily available regarding the hourly frequency distribution of rain occurrences in this area and for this season was inadequate to establish with certainty the time of minimum frequency. However, the frequency distribution of hourly occurrences for three stations in the area, the Weather Bureau Airport Station in Peoria, Peoria Lock and Dam, and Princeton, showed a minimum near 2000 CST during the period studied (1948–51). Assuming this indicated minimum is representative of the area, the use of 24-hour precipitation measurements ending at 1800 CST will result in a near minimum of unreal or apparent rain occurrences due to rain continuing into the next period. The total rainfall for each station on each day was tabulated for the four summer (June, July, and August) seasons, 1949 through 1951.

This area covers about 4,000 square miles and observations from 20 stations were used to determine the distribution over the area. If a station density of one per 200 square miles is adequate to describe the distribution of showers, each individual shower must, on the average, cover at least this much area during its life cycle. Some evidence that this is true is shown by the total rainfall associated with several individual thunderstorms in Ohio [3] where the measurable rainfall extended over areas of from about 150 to 250 square miles. Additional evidence on the adequacy of this station density is shown in figure 1 where the number of stations used to represent an area (1, 5, 10, 15, or 20) is plotted against the percentage of days during which at least one station within the appropriate group reported rain. For example, if one station, Peoria, is used to represent the area, rain was reported on about 29 percent of all days. This compares with about 30 percent over a 47-year record. In using 5 stations to represent

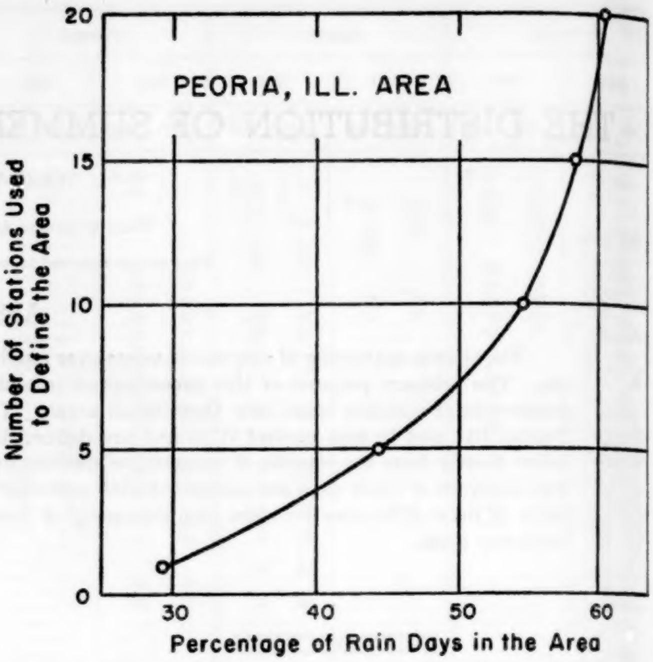

FIGURE 1.—Diagram showing the percentage of days when at least 1 station within a group of 1, 5, 10, 15, or 20 stations reported rain. Data are from the Peoria, Ill., area and are for the four summer seasons, 1948–51.

the area, it was found that at least 1 of these stations reported rain on 44 percent of all days. The station density was then increased to include these 5 stations plus 5 more to make 10, and then repeated for 15 and finally for all 20 stations. A smooth curve was fitted to these points by eye and it may be noted that between 15 and 20 stations the curve approaches independence of the abscissa. Thus, the precipitation data from these 20 stations provide a reasonably good indication of the precipitation coverage.

These data were then summarized to show the total number of days during which no station reported rain, any 1 station of the 20 reported rain, any 2 stations of the 20 reported rain, etc., through all 20 stations. A cumulative total, converted to percent of all days, was made, starting with rain at no station and on through 20 stations (100 percent). These data were plotted on figure 2 and a smooth curve (broken line) was fitted by eye. Only the curve itself is shown here in order that a comparison may be made between the various areas. It is interesting to note the difference between the curve based upon the data from 20 stations in the Atlanta area and that for the Peoria area. In the Peoria area there are about 10 percent more days when no station reported rain and about 3 percent more days when 20 stations reported rain. On only 54 percent of the days in the Peoria area is there some scattering (rain at at least 1 station but not all 20) of showers as compared with 67 percent of all days in the Atlanta area. This apparent difference of scattering between the two areas, amounting to 13 percent of all days, is largely accounted for by the difference in the frequency of rain in the two areas, rather than by any difference in the distribution of occurrences. That is, on only 29 percent of all days in the Atlanta area was no rain observed

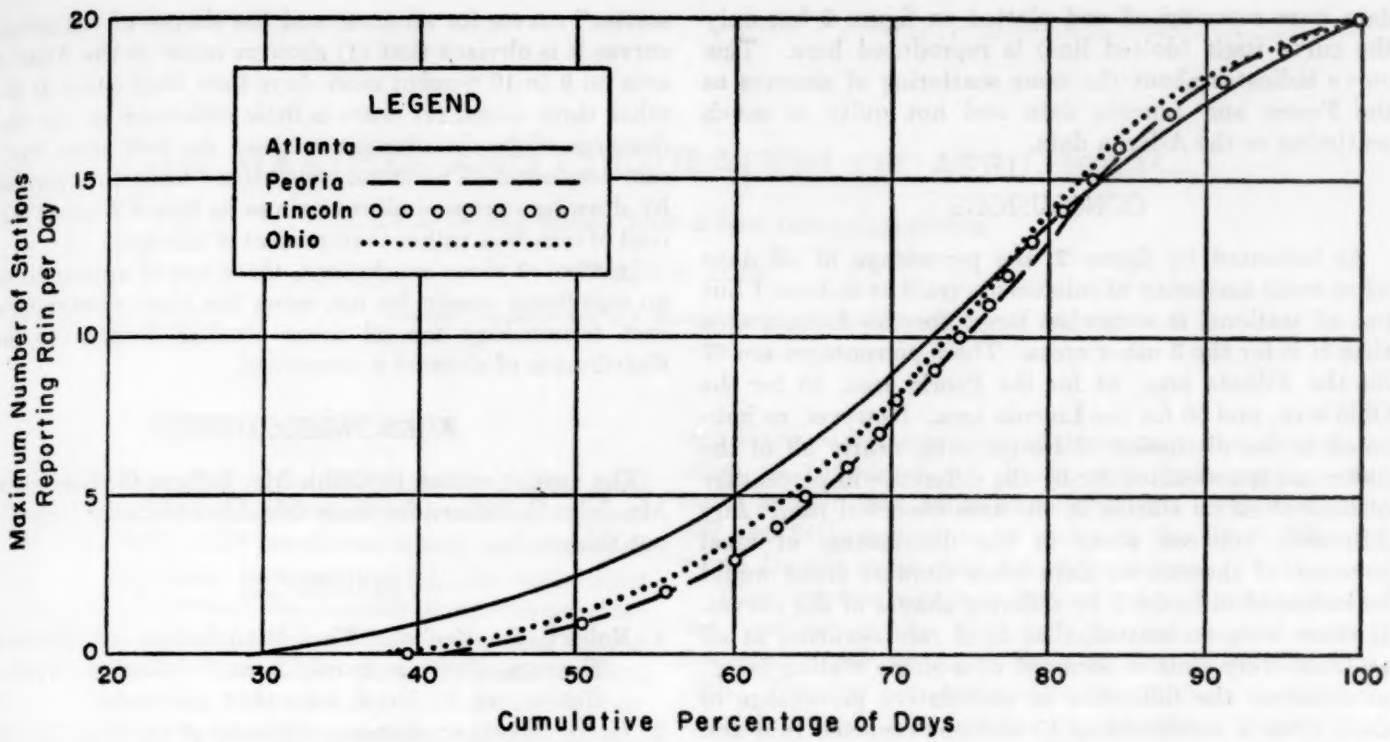

FIGURE 2.—Diagram showing the percentage of days when the maximum number of stations reporting rain was not greater than that indicated. Data from the Atlanta, Ga., and Lincoln, Nebr., areas are for the 6 summer seasons, 1946–51; data from the Peoria, Ill., area are for the 4 summer seasons, 1948–51; data for the east central Ohio area are for the 5 summer seasons, 1948–52.

as compared to 39 percent of all days in the Peoria area, which accounts for 10 percent of the 13 percent difference in apparent total scattering.

RESULTS FOR THE LINCOLN AREA

The diurnal frequency distribution of summer rainfall occurrences in the Lincoln area [4] shows a minimum shortly after noon with a gradual increase to a maximum soon after midnight. Available reports made it necessary to select stations for this study that reported around 1900 CST so that there will be more days during which rain was reported than there were rain periods. In addition, the two Weather Bureau stations at Lincoln reported 24-hour amounts ending at midnight (near maximum frequency of occurrence) so that a time lag is also introduced. These factors tend to indicate more scattering of showers than actually occurs.

Data were used for six summer seasons, 1946 through 1951, and they were summarized as in the case of the data for the Peoria area. A curve for the Lincoln data is not shown in figure 1, for these data gave a curve nearly identical to that for the Peoria area. Moreover, the Lincoln data for figure 2 are so similar to those for the Peoria area it was impracticable to show both curves in the same figure. Therefore, only the plotted points (crosses) for the Lincoln data are shown in figure 2. There is obviously no significant difference in the scattering of showers between this and the Peoria area.

RESULTS FOR THE EAST CENTRAL OHIO AREA

The hourly frequency of summer rain in central Ohio was found by Martin [5] to be at a minimum around 0400 and at a maximum around 1800 local time. Thus, of those available, the 0700 observations are the most desirable for this study while the midnight observations would introduce more apparent rain occurrences. It appeared impossible to find an area in Ohio where there was a good concentration of stations reporting at the same time of day, as was true of the Peoria area. The best concentration provided data from 20 stations with measurements around 0700 for three seasons (1950–52), and these 20 stations were selected for use in this study. In order to increase the amount of data, these 20 stations or substitutions were used for an additional two seasons (1948–49). During the 1949 season, substitute stations with measurements around 0700 were used on about one-third of the days; and during the 1948 season, substitute stations were used on about one-half of the days including three stations with midnight observations. In addition, it was necessary to substitute stations that were located a little outside the 35-mile radius in a few cases. These factors will introduce some error in the data and indicate scattering of showers that does not exist. However, these data at least provide some approximation of the areal coverage. The curve for this area is not shown in figure 1 because, as in the case of the Lincoln area, it is nearly identical to the curve for the Peoria area. The

data were summarized and plotted on figure 2 but only the curve itself (dotted line) is reproduced here. This curve indicates about the same scattering of showers as the Peoria and Lincoln data and not quite as much scattering as the Atlanta data.

CONCLUSIONS

As indicated by figure 2, the percentage of all days when some scattering of rain occurs (rain at at least 1 but not all stations) is somewhat larger for the Atlanta area than it is for the 3 other areas. These percentages are 67 for the Atlanta area, 54 for the Peoria area, 55 for the Ohio area, and 56 for the Lincoln area. However, as indicated in the discussion of Peoria data, nearly all of the difference is accounted for by the difference in percentage of time when no station in the area observed rain. Any difference between areas in the distribution of areal coverage of showers on days when showers occur would be indicated in figure 2 by differing shapes of the curves. If there were no scatter, that is, if rain occurred at all stations every time it occurred at a single station (e. g., at Atlanta) the difference in cumulative percentage of days when a maximum of 19 stations reported rain and when no station reported rain would be zero. In this case the Atlanta curve would be a straight line running along the 66 percent line from zero through 19 stations and then to the point for 100 percent and 20 stations. The "no-scatter" curve for the Peoria area would run along the 71 percent line, that for the Lincoln area along the 68 percent line, and for the Ohio area along the 68 percent line. The deviation of the actual curve from the curve for "no-scatter" is one measure of the amount of scattering, and any tendency for the actual curve to approach the shape of the "no-scatter" curve indicates deviation from perfectly random scattering. Considering the "no-scatter" curves for all areas and the shapes of the actual curves it is obvious that (1) showers occur in the Atlanta area on 9 to 10 percent more days than they occur in the other three areas, (2) there is little difference in the randomness of the distribution between the four areas when rain does occur. This latter conclusion was further verified by drawing curves similar to those in figure 2 using percent of rain days rather than percent of all days.

In view of these conclusions, there would appear to be no significant reason for not using the same shower forecast terminology for all areas studied, insofar as the distribution of showers is concerned.

ACKNOWLEDGMENT

The author wishes to thank Mr. Robert G. Beebe and Mr. John C. Ballard for their valuable assistance throughout this study.

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THE WEATHER AND CIRCULATION OF APRIL 1953[1]—

A Cold, Stormy Month With a Low Index Circulation

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WEATHER HIGHLIGHTS

The weather in most of the United States during April 1953 was unusually cold and stormy, in striking contrast to the abnormally mild conditions which had prevailed during the preceding four months [1]. In some cities (Denver and Cheyenne) the monthly mean temperature during April (Chart I-A) was actually lower than during March; while at others (Charleston, Tallahassee, Fort Worth) it was only a few tenths of a degree higher. Chart I-B shows that temperatures averaged below normal in all parts of the country except for eastern and southern border areas. The largest departure from normal, −7° F., was reported at both Glasgow, Mont. and Valentine, Nebr., while departures of −6° F. were observed at many stations in the Northern Plains. In Montana, Billings had its coldest April on record and Havre experienced below-normal temperatures on all but 6 days of the month.

The cold weather first affected the West Coast and then gradually spread eastward. During the first week of April nighttime temperatures in the State of Washington dropped to the lowest levels since November 1952. On the 7th the Napa Valley of California experienced its first major freeze since 1936. Record-low temperatures were established on the 8th at Winnemucca, Nev. (9° F.) and on the 9th at Fresno, Calif. (32° F.). The coldest week of the month east of the Continental Divide was the period from April 14–20, when temperatures averaged as much as 15° F. below normal in parts of the Northern Plains. North Dakota reported the coldest mid-April week on record, with temperatures as low as 0° F. on the 17th.

April's cold weather was accompanied by unusually heavy snowfall (Charts IV and V). In parts of the Northern Rockies and Northern Plains some snow fell in every week of the month. During the week ending April 13, 24 inches accumulated on the eastern edge of Great Salt Lake in Utah, while 10 to 13 inches fell in northwestern Kansas. On the 14th Boston had its heaviest snowfall this late in the spring season during the last 35 years. During the week ending April 20 snow was observed in many parts of the country including such cities as Cleveland, St. Louis, and Indianapolis. As much as 12 inches fell in parts of North Dakota on the 23d and 24th, and up to 14 inches in the Black Hills of South Dakota on the last day of the month.

Other weather elements besides cold and snow made the headlines during April. High winds caused severe duststorms over the entire State of New Mexico on the 17th. Freezing rain occurred in the southern sections of Kansas, Illinois, and Indiana on the 18th. During the last week of the month torrential rains, high winds, and floods caused considerable damage in the lower Mississippi Valley, where a new 12-hour precipitation record of 8.75 inches for Vicksburg, Miss., was established on the 29th. Thunderstorms and hail occurred at intervals during the month throughout the South. Perhaps the most spectacular feature of the weather was the unusual frequency of tornadoes. The total number of tornadoes reported in the United States during the month was 65, almost three times the normal number for April. Over a dozen tornadoes occurred in the State of Alabama alone, most of them during the disastrous weekend of the 18th–19th.

LOW INDEX CIRCULATION

The cold, stormy weather in the United States was produced by a hemispheric circulation of the classic low index type. According to Willett [2]

> the low index circulation pattern, in contrast to the high index pattern, is characterized by:
>
> 1. A relatively strong poleward temperature gradient, at least between sea level and the 3-km. level.
>
> 2. Relatively weak zonal westerlies at sea level which increase to relatively strong aloft, with a tendency to be displaced equatorward, that is, an intensified and expanded circumpolar vortex in the upper troposphere.
>
> 3. Strong polar easterlies as a result of a relatively strong sea-level polar anticyclone, which in turn is produced primarily

[1] See charts I-XV following page 129 for analyzed climatological data for the month.

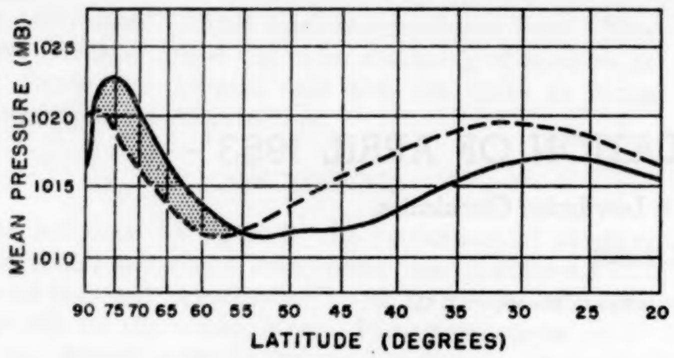

FIGURE 1.—Mean sea level pressure profile in the Western Hemisphere for March 30-April 28, 1953, with normal April profile dashed. Note positive anomaly of pressure north of 55° N. (shaded area) and negative anomaly to the south.

from a weakening of the subtropical high pressure belt.

4. In middle latitudes a relatively strong meridional circulation at sea level which tends to weaken with height as the zonal westerlies become stronger.

Most of this idealized description can be applied directly to the observed circulation of the Western Hemisphere during April 1953. At sea level the month was characterized by an excess of pressure in polar regions, north of 55° N., and a deficit of pressure to the south, at middle and low latitudes (fig. 1). As a result the polar easterly index was unusually high, while both the zonal westerlies and subtropical easterlies were abnormally weak (table 1). These low index features were particularly well marked

TABLE 1.—*Monthly mean indices in the Western Hemisphere during April 1953 (in meters per second)*

Index	Level	Observed	Normal	Departure from normal
Polar easterlies, 70°N.-55°N.	Sea level	3.0	1.6	+1.4
Zonal westerlies, 35°N.-55°N.	Sea level	1.2	2.5	−1.3
Subtropical easterlies, 35°N.-20°N.	Sea level	−0.1	1.8	−1.9
Polar westerlies, 55°N.-70°N.	700 mb	0.8	3.7	−2.9
Zonal westerlies, 35°N.-55°N.	700 mb	7.3	8.3	−1.0
Subtropical westerlies, 20°N-35°N.	700 mb	7.9	5.8	+2.1

in the Atlantic and North America, where monthly mean pressures in the polar anticyclones were as much as 12 mb. above normal, while the Azores-Bermuda high pressure belt was considerably weaker than normal (Chart XI and fig. 2). It is also noteworthy that the meridional circulation in middle latitudes was relatively strong, not only in the Atlantic and North America, but also throughout the Pacific.

At the 700-mb. level the April circulation also conformed closely to Willett's description of the low index state. Monthly mean heights were generally above normal in the polar region and at extremely low latitudes, while below-normal heights prevailed at most middle latitudes (fig. 3). As a result the subtropical westerlies were extremely strong, even stronger than the zonal westerlies, but the polar westerly index was well below normal (table 1). These features are well illustrated in the zonal wind speed profile for the Western Hemisphere (fig. 4) which shows that the westerlies were generally weaker than normal

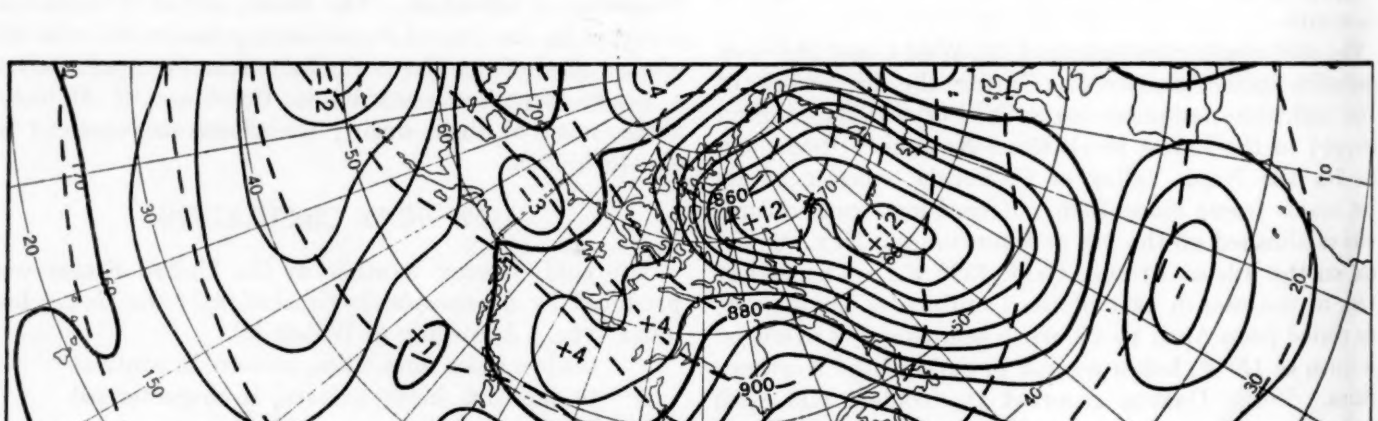

FIGURE 2.—Mean sea level pressure departure from normal for March 30-April 28, 1953 (at intervals of 2 mb.). Dashed lines indicate normal thickness (700 mb.-1,000 mb.) for April (in tens of feet). Note anomalous components of flow into the United States from cold source regions in Hudson Bay and eastern Pacific.

north of 43° N. but stronger than normal south of this latitude. Furthermore, the strongest westerlies, around 33° N., were about 5° south of their normal latitude and almost 2 meters per second stronger than normal. The regional distribution of total horizontal wind speed at 700 mb. is delineated in figure 5a. This map shows that the axis of strongest wind speed was located around 40° N. in the eastern Pacific and 5° to 10° farther south in the Atlantic. In North America this "jet stream" was split into two parts, with the principal branch passing through Mexico and Texas and a weaker branch traversing the northwestern United States. Figure 5b simplifies this picture somewhat since it shows that wind speeds were generally weaker than normal in Canada but stronger than normal in the United States, so that the westerlies showed a distinct tendency "to be displaced equatorward." Thus the April low index circulation was clearly characterized by an "intensified and expanded circumpolar vortex" at least at the 700-mb. level.

The monthly mean 200-mb. chart (fig. 6) is also indicative of the low index state since it shows a pronounced jet stream at relatively low latitudes (25°–40° N.) girdling

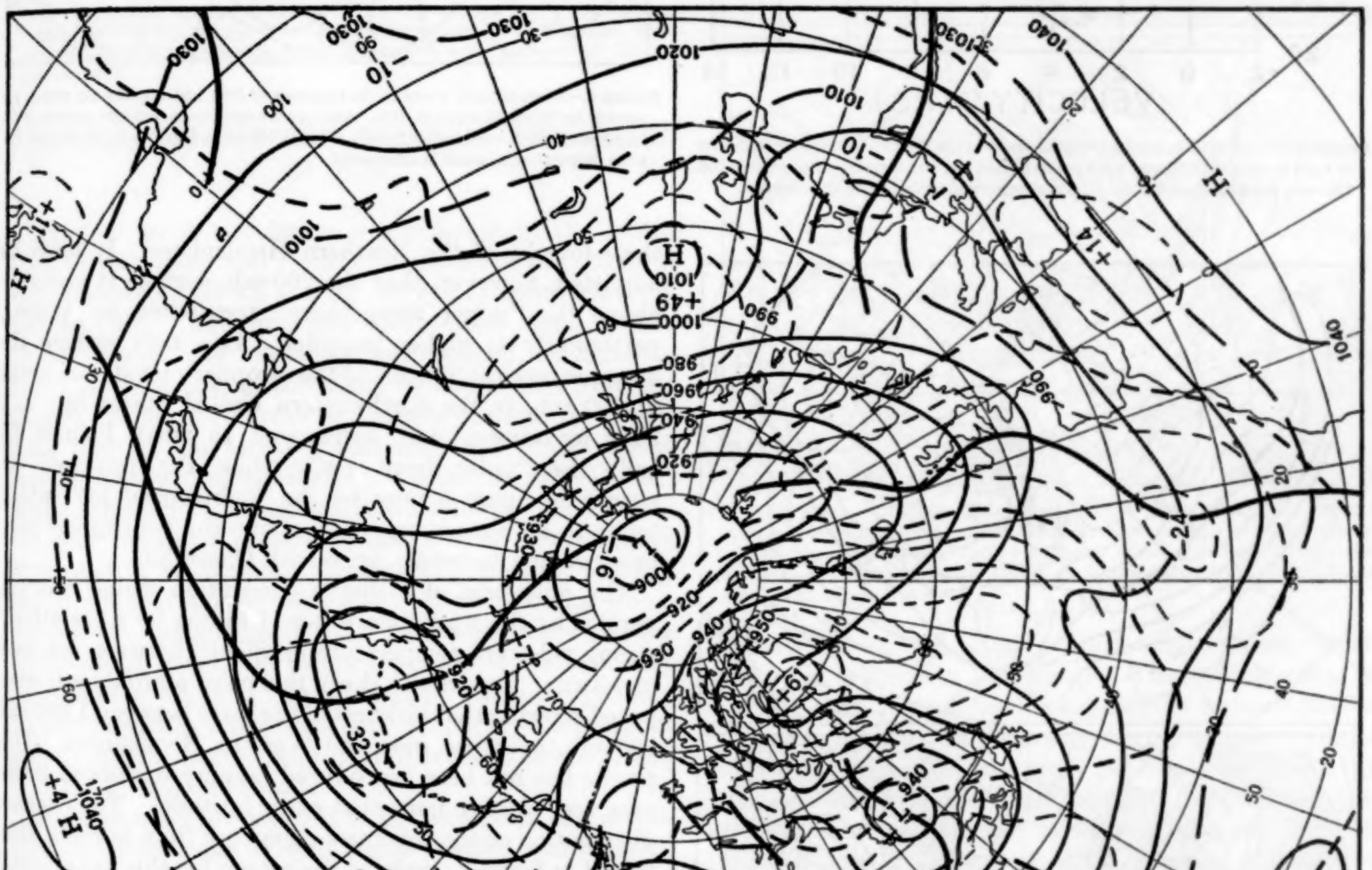

FIGURE 3.—Mean 700-mb. height contours and departures from normal (both in tens of feet) for March 30–April 28, 1953. Features of interest include strong blocking ridge over Baffin Bay, subpolar Low displaced south to James Bay, and negative height anomalies throughout United States.

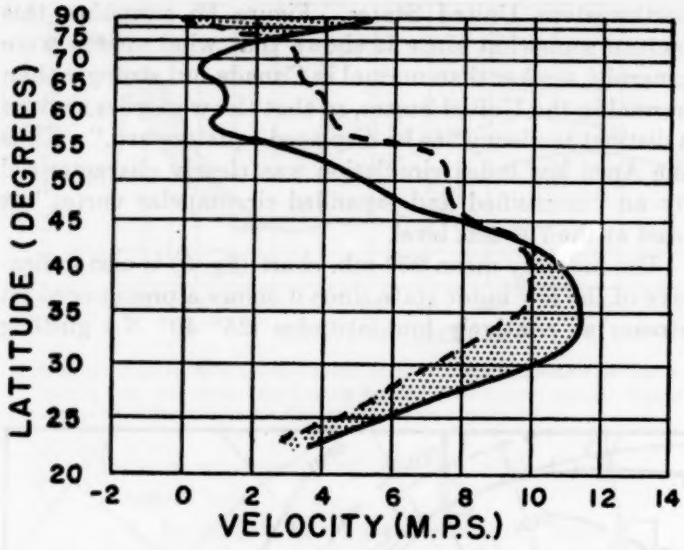

FIGURE 4.—Mean 700-mb. zonal wind speed profile in the Western Hemisphere for March 30-April 28, 1953, with normal April profile dashed and area of positive anomaly shaded. The west wind maximum at 33° N. was stronger and farther south than normal.

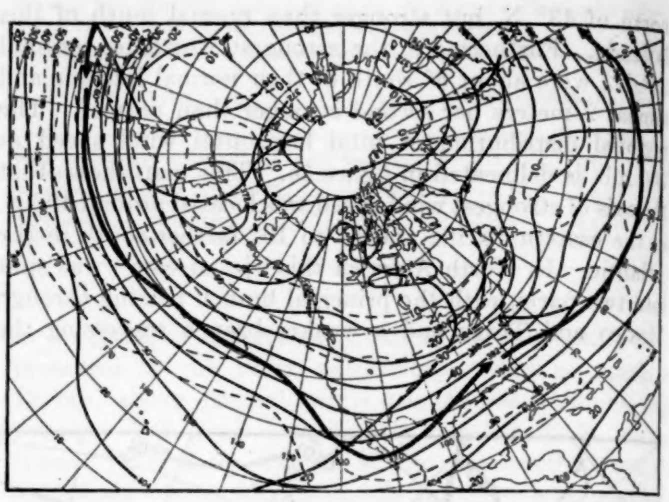

FIGURE 6.—Mean 200-mb. contours (in hundreds of feet) and isotachs (in meters per second) for March 30-April 28, 1953. Solid arrows indicate the average position of the jet stream, which was located directly over the 700-mb. jet stream in the Pacific, but a few degrees farther south in other areas.

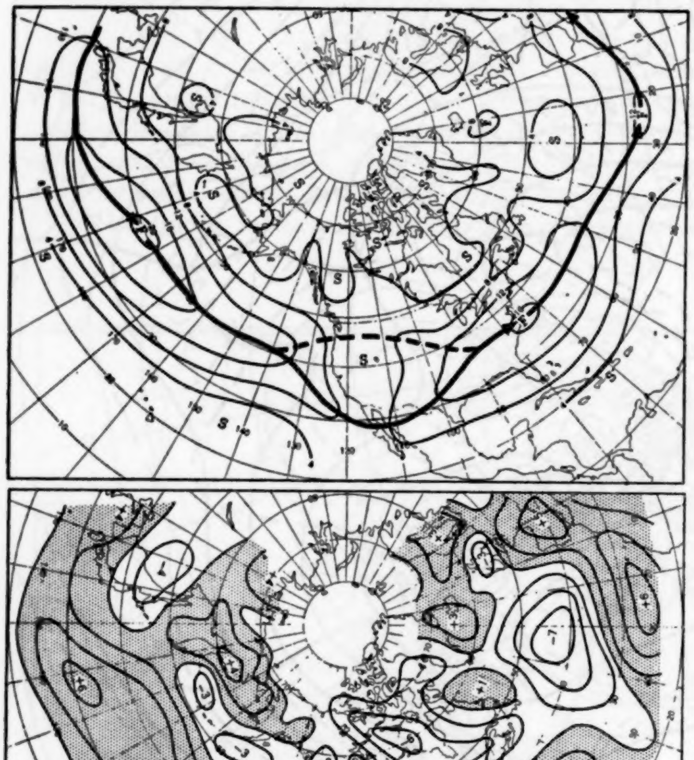

FIGURE 5.—Mean 700-mb. isotachs (*a*) and departure from normal wind speed (*b*) (both in meters per second) for March 30-April 28, 1953. Solid arrows indicate the average position of the jet stream, which was south of its normal location in nearly all sectors. Dashed line delineates secondary zone of maximum wind speed across northwestern United States.

three-fourths of the Northern Hemisphere. It must be admitted, however, that the 200-mb. surface is probably above the "upper troposphere" mentioned by Willett, particularly at higher latitudes. This may explain the disappearance at 200 mb. of the secondary jet stream noted at 700 mb. in the northwestern United States (fig. 5a). This jet stream may correspond to what Palmén [3] calls the "polar front jet", while the principal jet stream in figure 6 may be the "subtropical jet" which tends to lie almost vertically above the subtropical high pressure belt (compare fig. 6 with Chart XI).

The only part of Willett's description which was not verified by the April data is item 1 calling for a "relatively strong poleward temperature gradient" between sea level and 3 km. In order to check this point a profile was computed showing the thickness of the layer between 1,000 and 700 mb. averaged over the Western Hemisphere. This profile has not been reproduced because it showed thickness very close to normal at every latitude. The regional distribution of the thickness departure from normal, illustrated in figure 7, explains the reason for this near normal profile. This chart shows that the poleward temperature gradient at middle latitudes was indeed stronger than normal in the eastern Pacific and western North America, but it was definitely weaker than normal in the western Atlantic and eastern North America. However, it was precisely the Atlantic sector which had the lowest index conditions, whereas the Pacific circulation was more of the high index type. For this reason, and because of the well-known fact that blocking Highs are usually warm (e. g., the large High over Greenland and northeast Canada this month) while Lows at low latitudes are frequently cold, the first item of Willett's description does not appear to be typical of the low index state, at least not in the Western Hemisphere.

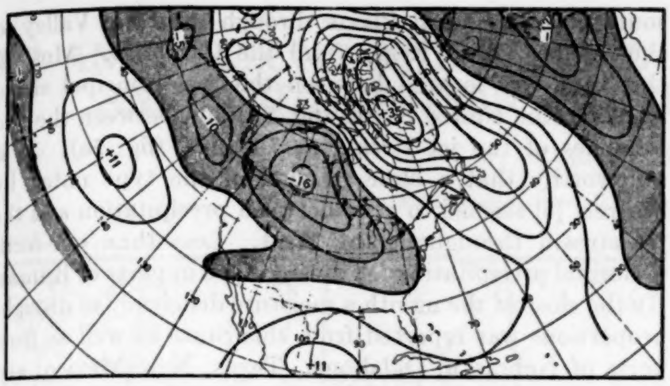

FIGURE 7.—Mean departure from normal of thickness (700 mb.–1,000 mb.) for March 30–April 28, 1953, analyzed for intervals of 50 feet with centers labeled in tens of feet. Below normal thicknesses (shaded) covered most of the United States, with center of −160 ft. corresponding to mean virtual temperature about 5° C. below normal. Note warmth of blocking High over northeast Canada with center almost 12° C. (350 ft.) above normal.

INTERRELATION BETWEEN WEATHER AND CIRCULATION

The low index circulation was reflected in the southward displacement and meridional nature of the tracks of anticyclones (Chart IX and fig. 8a) and cyclones (Chart X and fig. 8b). The principal anticyclone track in the United States was extremely well defined. Migratory Highs entering the northwestern United States from the eastern Pacific and northwest Canada plunged far southward through the Great Plains and Gulf States before moving eastward at unusually low latitudes (30°–35° N.) in the Atlantic. Cyclone tracks were not as concentrated but they showed a definite tendency to be located south of normal in all sectors. Considerable stalling, looping, and meridional motion were also evident. Storms which entered the United States from the Pacific traversed the northwestern part of the country, parallel to the secondary jet stream at 700 mb. To the south, where the primary jet was located, cyclogenesis and "secondary" formations were frequent, particularly near Nevada, Colorado, and the Middle Atlantic Coast.

The unusually low latitude of the jet stream and associated anticyclone and cyclone tracks enabled cold polar air to penetrate most of the United States. This cold air came from two principal sources, Hudson Bay and the eastern Pacific, and followed approximately the same trajectory as the principal anticyclone track illustrated in figure 8a.[2] In April these two regions are normally among the coldest parts of North America and vicinity; this year the anomalous components of flow, both at sea level and aloft, came directly from these cold sources into the United States. In order to illustrate this, the normal thickness lines for April (1,000–700 mb.) have been superimposed on the departure from normal of the monthly mean sea level pressure (fig. 2). The advection of cold

[2] For an example of a severe cold outbreak associated with a polar High originating just west of Hudson Bay, see adjoining article by Malkin and Holzworth.

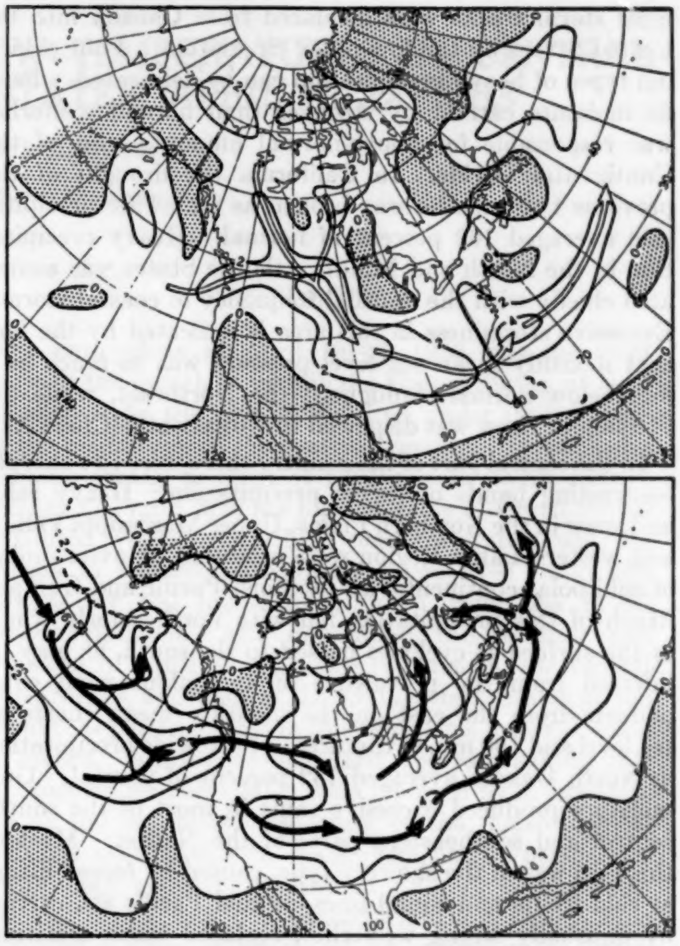

FIGURE 8.—Frequency of anticyclone passages (a) and cyclone passages (b) (within 5° squares at 45° N.) during April 1953. Well defined anticyclone tracks are indicated by open arrows and cyclone tracks by solid arrows. Effect of the low index is apparent in meridional components and southward displacement of tracks. (All data derived from Charts IX and X.)

"climate" from the Hudson Bay and eastern Pacific sources is evident in this figure. The cooling effect of stronger than normal west winds in the western part of the country this month contrasts sharply with its warming effect during the past winter [1]. A ready explanation is provided by the normal thickness lines, which show that air over the Pacific ocean is warmer than air over the United States in winter but colder in spring and summer [4].

After entering the United States the cold air was prevented from rapid warming by excessive cloudiness (Charts VI and VII), abundant storminess (Chart X), and cyclonic circulation with below normal heights at 700 mb. (fig. 3). The only places sheltered from the recurrent cold outbreaks were the North and Middle Atlantic Coasts, the Rio Grande Valley, and parts of Florida and the Gulf Coast. In those regions monthly mean temperatures averaged a few degrees above normal under the influence of some southerly components of flow east of the primary trough at 700 mb.

Total precipitation during the month (Chart II) generally exceeded normal amounts (Chart III) as the prin-

cipal storm tracks were displaced from Canada into the United States by the low index circulation. Four principal types of heavy precipitation can be delineated. Pacific moisture carried by stronger than normal westerlies was responsible for above-normal amounts west of the Continental Divide. In California the drought of the previous two months was broken as statewide precipitation averaged 162 percent of normal. Heavy precipitation in the North and Middle Atlantic States was associated chiefly with the unusual frequency of coastal storms. Excessive storminess in this area is indicated by the fact that monthly mean sea level pressure was as much as 8 mb. below normal throughout the Northeast, while the "Icelandic" Low was displaced to Nova Scotia (Chart XI).

In the central part of the country there were two sharply contrasting bands of excess precipitation. Heavy rains and snow in the Northern Plains, Upper Mississippi Valley, and western Lakes Region were produced by overrunning of cold polar continental air by moist Pacific and Gulf air. Much of this precipitation fell with northeasterly winds at the surface as cyclones passed to the south, as may be inferred from the prevalence of anomalous wind components from the east on the monthly mean charts at sea level and 700 mb. in this area. Statewide precipitation in North Dakota averaged 182 percent of normal. Gulf moisture produced excessive rains in most of the south-central and southeastern parts of the Nation. Most of this rain was of the showery type, caused by forced lifting at cold fronts and squall lines, as cold Pacific air, carried by unusually strong westerlies, replaced warm maritime tropical air in the area. The resulting instability was sufficiently great to produce numerous thunderstorms and tornadoes throughout the South.

In between these two bands of heavy precipitation a narrow zone of subnormal rainfall extended from the southern and central Plains through the Ohio Valley to the eastern Lakes Region and the Carolinas. Most of this zone was located between the two principal storm tracks in the United States (fig. 8b) and between the two branches of the jet stream at 700 mb. (fig. 5a). It is noteworthy that a close relation of the type noted by Starrett [5] seemed to exist between precipitation and the jet stream throughout the West. Less than one-forth of normal precipitation fell during April in parts of Kansas. By the close of the month a moisture deficiency of drought proportions was reported from this State as well as from parts of Nebraska, Oklahoma, Texas, New Mexico, and Colorado. Light rainfall in this area was also associated with downslope motion in stronger than normal northwesterly flow to the rear of the 700-mb. trough.

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THE ANTICYCLONE AND RECORD LOW TEMPERATURES IN CENTRAL AND SOUTHEASTERN UNITED STATES, APRIL 19–22, 1953

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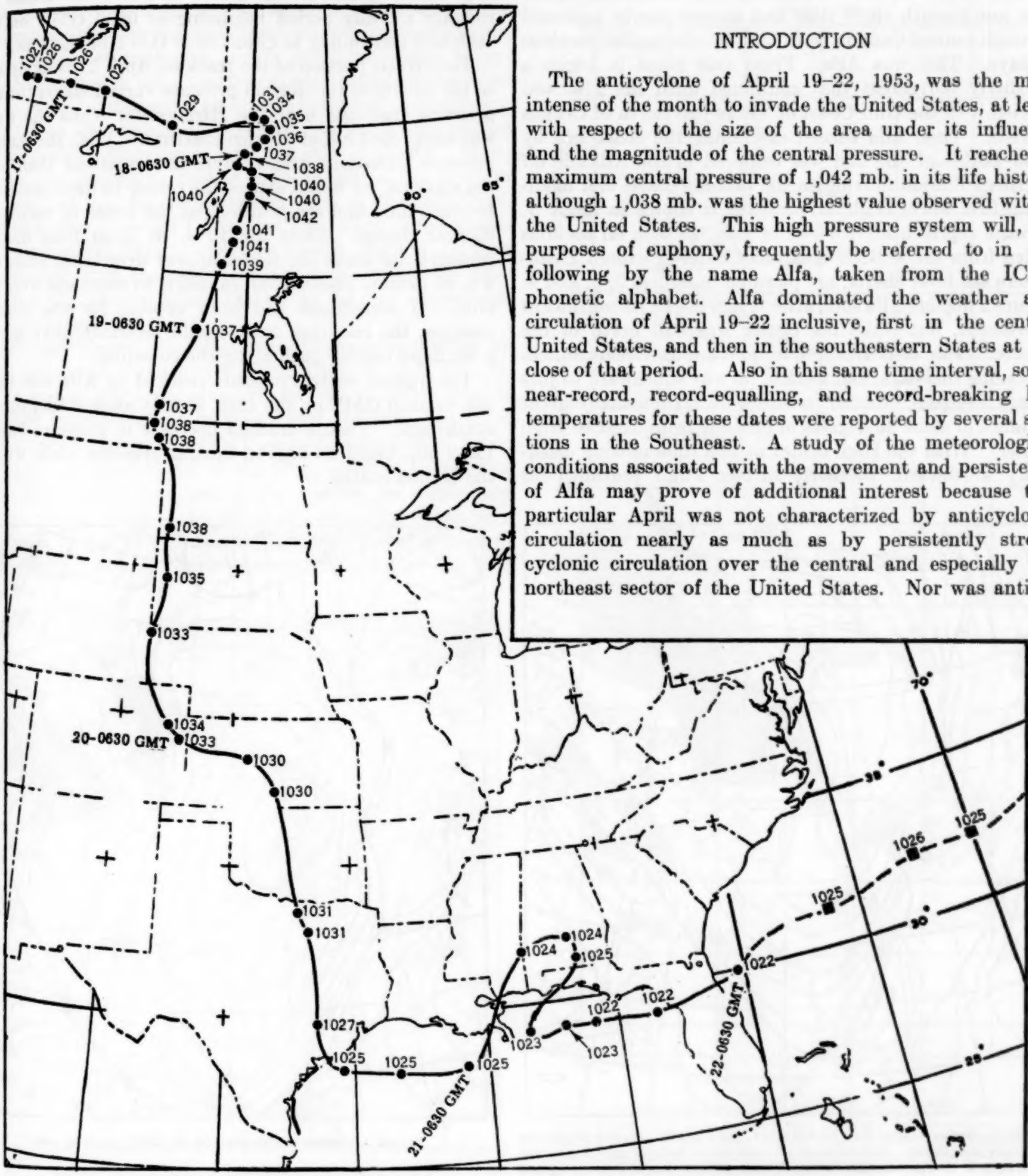

FIGURE 1.—The path followed by the center of Alfa. Central pressures and positions for every 3 hours are shown as dots, except over the Atlantic Ocean where positions are for every 6 hours and are indicated by squares. Reference dates and times are given for the 0630 GMT positions.

INTRODUCTION

The anticyclone of April 19–22, 1953, was the most intense of the month to invade the United States, at least with respect to the size of the area under its influence and the magnitude of the central pressure. It reached a maximum central pressure of 1,042 mb. in its life history although 1,038 mb. was the highest value observed within the United States. This high pressure system will, for purposes of euphony, frequently be referred to in the following by the name Alfa, taken from the ICAO phonetic alphabet. Alfa dominated the weather and circulation of April 19–22 inclusive, first in the central United States, and then in the southeastern States at the close of that period. Also in this same time interval, some near-record, record-equalling, and record-breaking low temperatures for these dates were reported by several stations in the Southeast. A study of the meteorological conditions associated with the movement and persistence of Alfa may prove of additional interest because this particular April was not characterized by anticyclonic circulation nearly as much as by persistently strong cyclonic circulation over the central and especially the northeast sector of the United States. Nor was anticy-

clonic activity in evidence over the West and Southwest, where flat pressure gradients prevailed for most of the month, with no formation of a Great Basin High in this period.

PATH OF CENTER

At 1830 GMT, April 17 a high center of 1,034 mb. was located at 60° N., 100° W., having evolved from an extensive north-south ridge that had moved slowly eastward through central Canada from 115° W. during the previous 2 days. This was Alfa. From this point it began a southerly movement that continued until the 21st and carried it to the Gulf Coast of Texas just north of Corpus Christi. Then Alfa turned east along the coast and by 0630 GMT of the 22d was centered in the Atlantic off northern Florida moving to the east-northeast and modifying to a warm High. This path is shown in figure 1.

Some representative views of Alfa, as seen on sections taken from the WBAN Analysis Center Northern Hemisphere sea level charts, are pictured in figures 2, 3, and 4. Figure 2 depicts Alfa soon after it had begun its southward movement, but while the center was still north of the border. The cold front, due to various circumstances preceding this date, had worked its way southward to just below Tampico, Mexico, resulting in an unusually great distance of some 30 degrees of latitude from front to High center. With the High center at this time moving essentially southward, northerly surface winds prevailed for many hours over the southern States behind the cold front, contributing materially to the lowering of surface temperatures. Thus we note (figs. 2 and 3) that along a line from Minnesota to southeast Texas, and over an area several hundred miles on either side of this line, the surface and gradient winds were predominantly northerly. This extensive field of northerly winds prevailed over most of the United States to the south of the Great Lakes for roughly a 2-day period beginning at 0630 GMT on the 19th and continuing to about 0630 GMT of the 21st.

The erratic portion of the track on April 21 is likely due to the effects of the diurnal pressure variations on the flat pressure gradients near the High center while the track was along the Gulf of Mexico coastline. With the diurnal pressure variations greater over land stations than over sea stations, we would expect the center to have oscillated between land and sea positions at the times of maximum diurnal change. Thus the 10 a. m. local time diurnal pressure rise made the center appear over land, while the 4 p. m. diurnal pressure fall caused it to move out over the Gulf. If corrections had been applied for the diurnal changes, the resulting track would probably have shown a far more regular path along the coastline.

The highest central pressure reached by Alfa was 1,042 mb. at 1830 GMT on the 18th, shortly after it had turned southward. Then a gradual decrease in pressure began, 1,038 mb. being its highest central pressure while within the United States.

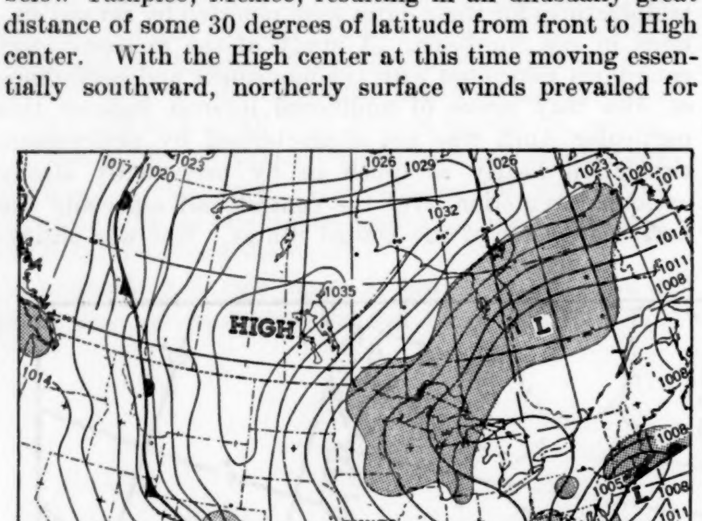

FIGURE 2.—Surface weather chart for 0630 GMT, April 19, 1953. Shading shows areas of active precipitation. Note northerly flow over eastern half of country on this map and figure 4. Lowest temperatures were experienced in the Southeast on the nights of April 19–22.

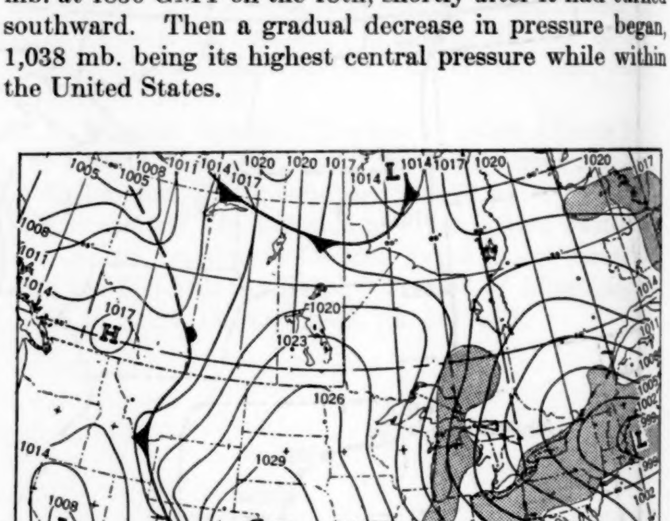

FIGURE 3.—Surface weather chart for 0630 GMT, April 20, 1953.

VERTICAL STRUCTURE

Of the Canadian stations making radiosonde observations regularly, Churchill was closest to the high center in the period from 1500 GMT of the 17th to 0300 GMT of the 19th; it was sufficiently close that these soundings may be used *per se* to represent the vertical structure through the center of Alfa at these times. The soundings are shown as a composite in figure 5. In each of the soundings observe the weak inversion off the ground and, above this, an extensive, nearly isothermal layer to approximately 750 mb., which is characteristic of the lapse rate in cold, continental air masses near their source region [1]. The corresponding curves for dew point temperature were omitted from figure 5 for the sake of clarity. However, the respective mixing ratio values in the stratum below 750 mb. for all points on each of the soundings, ranged closely around 1.3 gm. per kg. Each of the soundings in figure 5, when plotted on a Rossby diagram and then compared with the characteristic slopes given by Showalter [2], clearly labels the air as having continental Arctic properties.

Further insight into Alfa's vertical structure is given by the time cross section over Rapid City, S. Dak. (fig. 6). The warm, low tropopause of the typical Arctic type is clearly in evidence at 0300 GMT of the 19th, but fades away rapidly after 1500 GMT upon the approach of Alfa's center. At 0300 GMT of the 20th, about 6 hours after the High center had passed the station, no tropopause point can be found to the top of this sounding, which only reached to 200 mb.; however, the existence of a tropical type tropopause is now suggested. The top of the cold dome was very flat throughout the 24 hours of the time section and was located very near the 700-mb. level.

While Alfa satisfied most of the specifications given by Wexler [3] to be classified as a cold or polar type anticyclone of North America throughout the time interval of this study, comments are in order on the weakness of its

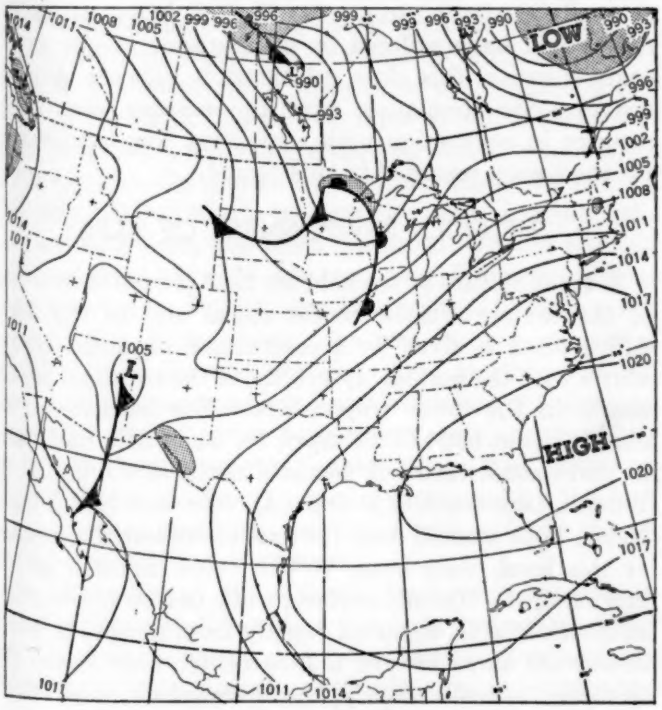

FIGURE 4.—Surface weather chart for 0630 GMT, April 22, 1953.

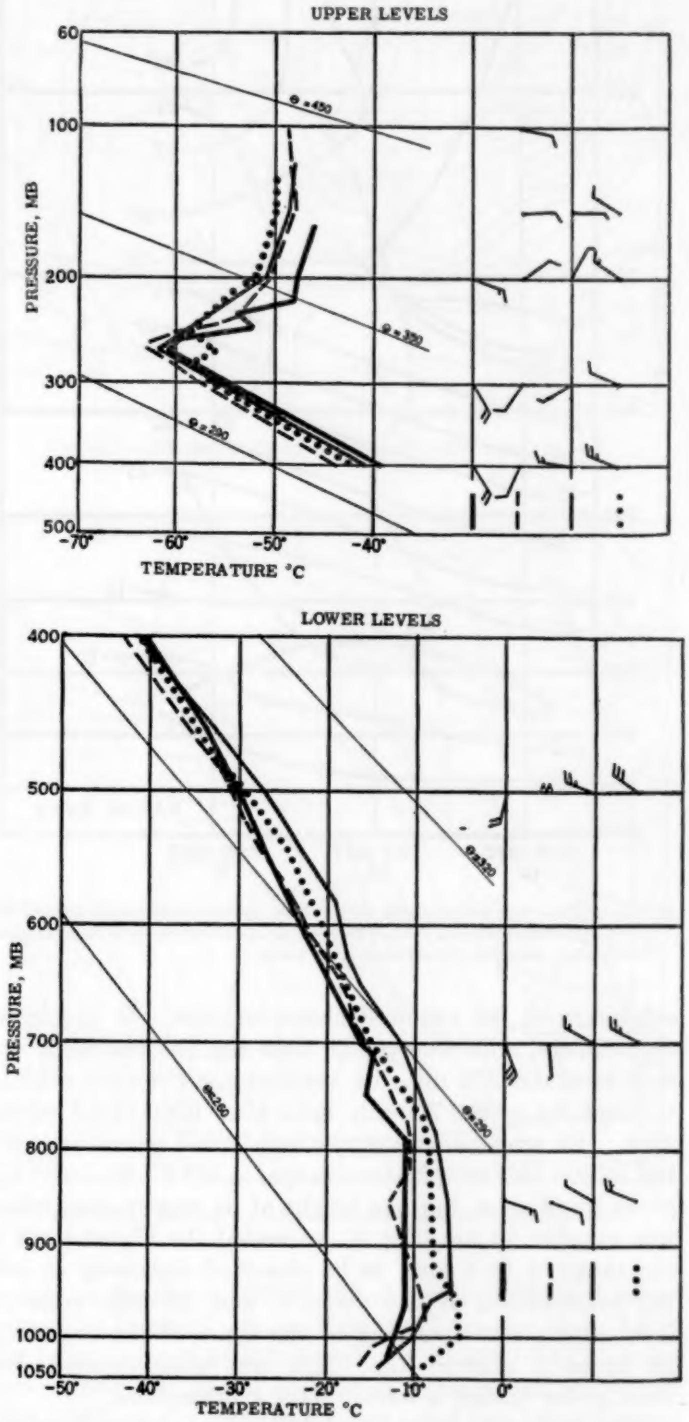

FIGURE 5.—Upper air soundings over Churchill, Manitoba at 1500 GMT, April 17 (heavy solid line), 0300 GMT, April 18 (dashed line), 1500 GMT, April 18 (thin solid line), and 0300 GMT, April 19 (dotted line). The winds are for the constant pressure levels at which they are plotted and correspond to the soundings as indicated by the key at the bottom right. A full barb indicates a speed of 10 knots.

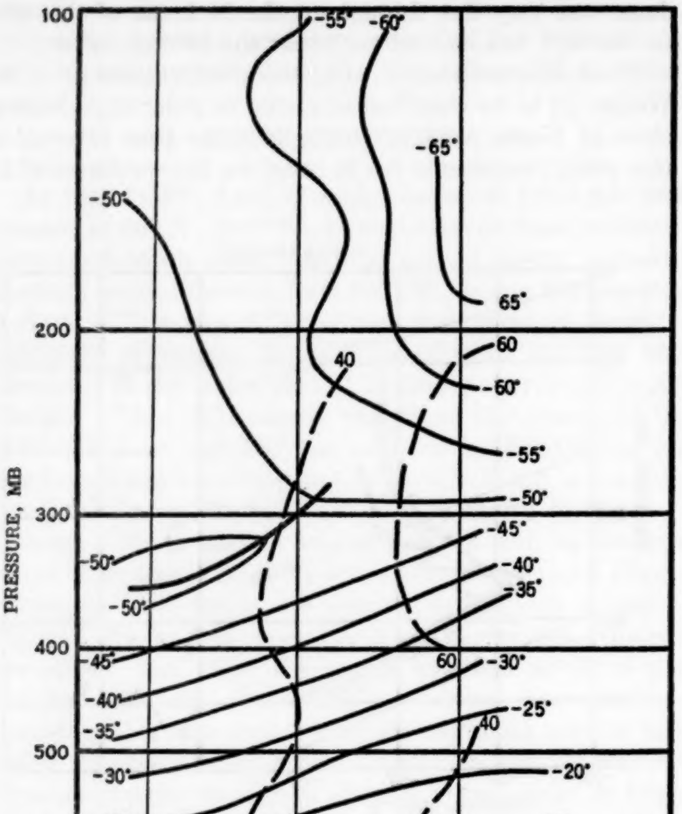

FIGURE 6.—Time cross section, Rapid City, S. Dak., showing soundings at 0300 and 1500 GMT, April 19 and 0300 GMT, April 20. Isotherms (solid lines) are in degrees Celsius (Centigrade). Isotachs (dashed lines) are in knots.

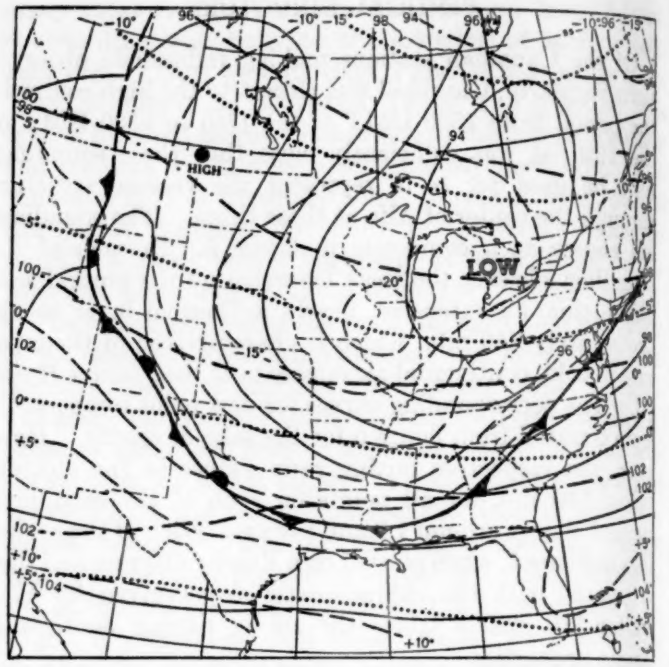

FIGURE 7.—700-mb. chart at 1500 GMT, April 19, 1953 superimposed on the normal April chart. Contours are in hundreds of geopotential feet; April 19 contours are given by solid lines and normal April contours by dash-dot lines. Isotherms are in degrees Celsius (Centigrade); April 19 isotherms are given by dashed lines and the normal April isotherms by dotted lines. The large dot over the word "High" shows the sea level position of Alfa.

exhibition of the required characteristics. In the lower troposphere, Alfa was colder than the environment, but only to about 850 mb.; its anticyclonic circulation failed to reach up to the 700-mb. level after 0300 GMT of the 20th. Its tropopause may be considered warm as it did fall within the temperature range, −50° C. to −65° C., given by Wexler, but the height of its tropopause, which was roughly 10 km. (fig. 5), exceeded the higher limit of the range, 5 to 8 km., to be classified distinctly as low. An examination of the 200-mb. and 150-mb. constant level charts corresponding to the times of the soundings in figure 5 showed that Alfa was distinguished, but weakly, for having a warm, lower stratosphere.

The apparent weakness of Alfa as a polar anticyclone, by the above criteria, was likely due at least partially to its immediate proximity to the quasi-stationary cyclonic circulation over the Great Lakes region. This cyclonic circulation was somewhat of the cold Low type, extending over a larger area than usual, and exhibiting colder temperatures than the environment at all constant levels in the troposphere up to and including the 500-mb. level, and was itself associated with a low, warm, polar type tropopause. On coming into close contact with such an environment already possessed of many of the properties generally attributed to a polar High, it is reasonable that Alfa would have suffered by comparison. Later we will again mention this close relationship between Alfa and the cyclonic circulation over the eastern part of the country in connection with some very low temperatures for the season recorded in this period.

THE COLD ENVIRONMENT OF ALFA

A vivid picture of the cold air that prevailed over most of the eastern United States ahead and to the left of Alfa's track is given by a comparison of actual temperatures with the normal [4] on some of the constant pressure charts in the lower troposphere. For instance, it was found that at 0300 GMT April 19, when Alfa had started to move southward but was still north of the border, the 700-mb. temperatures at Omaha, Nebr. and North Platte, Nebr., both located near the axis of coldest temperatures at this level, were some 15° C. below normal, and the heights of the 700-mb. surface, while not far below normal at North Platte, departed rapidly from normal in a steep downward slope toward a Low center near North Bay, Ontario. Simultaneously, the observed winds at 700 mb. over North Platte, Omaha, and stations in adjacent States to the north and east, had northerly components

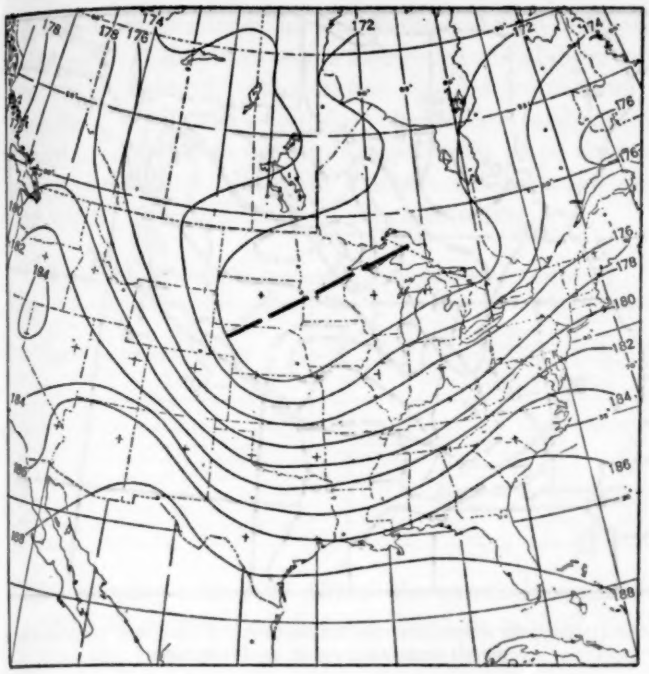

FIGURE 8.—1,000–500 mb. thickness chart at 0300 GMT, April 19, 1953. Thicknesses are in hundreds of geopotential feet. The heavy dashed line shows the axis of cold air. Note in figures 9 and 10 how this cold air axis moved east-southeastward ahead of the center.

distinctly greater than normal. This axis of low temperatures at 700 mb. moved steadily eastward and by 0300 GMT, April 21, was located along the Atlantic Coast. At this time the 700-mb. temperature at Cape Hatteras, N. C., for example, was 14° C. below normal and the height of the 700-mb. surface much lower than normal, with the wind having a slight northerly component as contrasted with the normal due west winds. A particular instance in this period, illustrating these departures from normal at the 700-mb. level, is shown in figure 7.

One of the outstanding features of the 1,000–500 mb. thickness charts in this period, three of which are reproduced in figures 8, 9, and 10, is the axis of cold temperature in this stratum, designated by a dashed line in the figures. This axis of low mean temperature moved east-southeastward with time, and when compared with the track of Alfa, is seen to have led the center at all times. A case study reported by Sutcliffe and Forsdyke [5] likewise made mention of a High center that was located west of a 1,000–500 mb. thermal trough. This condition is not unusual, but is mentioned as being indicative of the strong baroclinity of the lower atmosphere east of the High center. In superimposing the track of the High center on these thickness charts, we are also impressed by the fact that the center itself was associated with a fairly strong thermal gradient in this stratum.

The anomalies at the 500-mb. level were apparently similar to those at 700 mb. in this time interval, judging from the height departures from normal, temperature normals not being directly available. Some representative anomalies at the 500-mb. level were: 700 feet below

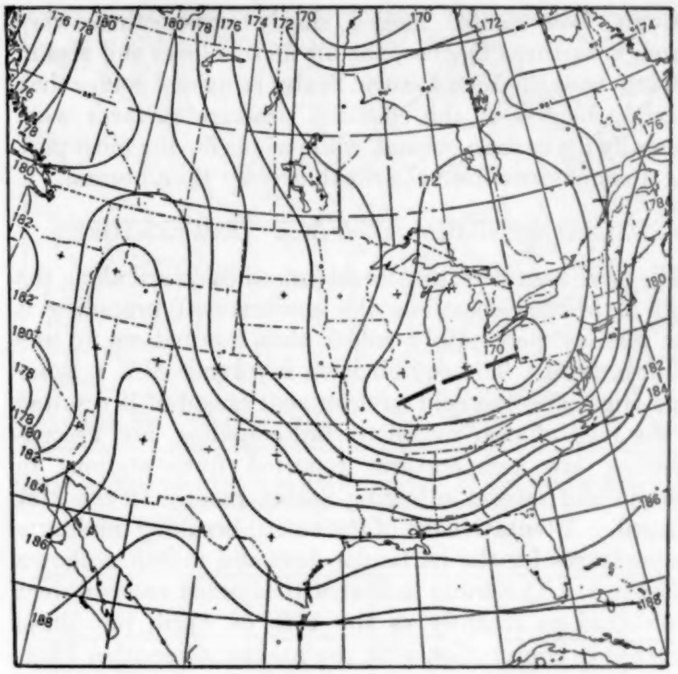

FIGURE 9.—1,000–500 mb. thickness chart for 0300 GMT, April 20, 1953.

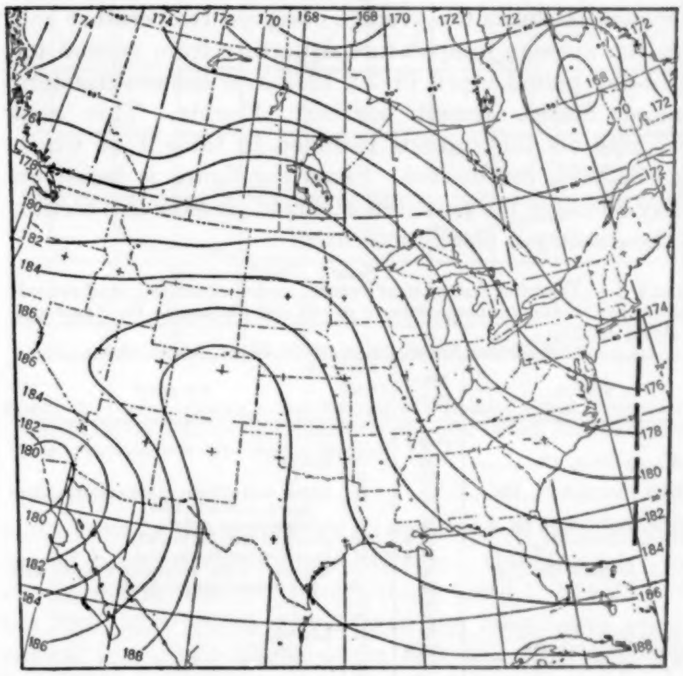

FIGURE 10.—1,000–500 mb. thickness chart for 0300 GMT, April 22, 1953.

normal at Omaha at 0300 GMT, April 19; 1,000 feet below normal at Rantoul, Ill. at 1500 GMT on the 19th; 1,000 feet below normal at Pittsburgh, Pa. at 0300 GMT on the 20th; 1,000 feet below normal at Washington, D. C. at 1500 GMT on the 20th; and about 1,000 feet below normal at both Mount Clemens, Mich. and Nantucket, R. I. at 0300 GMT on the 21st.

It is therefore apparent that in this period, as the

700-mb. Low center moved slowly east-northeastward through southern Quebec, the air at this level and also at 500 mb. over the East Central States remained colder than normal, heights of the constant pressure surfaces were generally lower than normal, while winds for the most part had a component from the north greater than normal.

RECORD MINIMUM SURFACE TEMPERATURES

We have seen how, as Alfa advanced eastward along the Gulf of Mexico coastline, the environment preceding it was already perceptibly colder than normal up to the 500-mb. level. The combination involving Alfa, a polar type High, and the cold environment ahead of it resulted in the rash of near-record, record-equalling, and record-breaking low temperatures reported from stations in Georgia and several adjoining States, mostly on the 21st of April. A tentative list of the record-breaking minimum temperatures for the particular date and month is shown in table 1. The table indicates that some records were established as recently as the 17th of April, but these occurred in conjunction with the passage of another High through the Southeast ahead of Alfa. By far the greater number of records were set with the passage of Alfa on the 21st.

Figure 11 shows the average observed temperature and also the average temperature departure from normal for the 4-day period, April 19–22, 1953 over the southeastern United States, except southern Florida. This area embraces all the stations included in table 1 at which records were established. Since the figure is based on 4-day averages the persistence of cold air over this section of the country is clearly indicated.

Table 1.—*Tentative list of near record, record-equalling, and record-breaking minimum temperatures at selected stations in the Southeast, April 1953*

Station	Min. temp.	Date	Remarks*
	(° F.)		
WBO, Anniston, Ala	29	21	Lowest.
	36	22	Do.
WBAS, Birmingham, Ala	31	21	Latest date of freezing temperatures since records began in 1896.
WBAS, Tallahassee, Fla	39	21	1 degree above record.
WBAS, Atlanta, Ga	36.6	20	Previous low 37.6 in 1901.
	33.4	21	Previous low 36.4 in 1914.
WBAS, Augusta, Ga	33	17	Lowest. Fifth consecutive April with sub-normal temperatures.
	34	20	
	30	21	Do.
	35	22	Do.
WBAS, Rome, Ga	27	21	Lowest.
WBAS, Savannah, Ga	36.3	21	Do.
WBAS, Spartanburg, S. C	35	20	
	34	21	
	36	22	
WBAS, Columbia, S. C	32	20	Do.
	29	21	Do.
WBAS, Charleston, S. C	34	21	Do.
WBO, Charleston, S. C	42	21	Lowest for so late in the season since 1910.
WBAS, Greenville, S. C	32	21	Latest freeze since records began in 1917.
WBO, Richmond, Va	33	21	
	34	22	
WBAS, Chattanooga, Tenn	29	21	Lowest (records began 1879).
WBAS, Nashville, Tenn	30.6	19	Lowest for this date.
WBAS, Wilmington, N. C	35	21	Lowest.
WBO, Raleigh, N. C	34	21	Record.
WBAS, Greensboro, N. C	29	17	Do.
	31	20	Do.
	31	21	Do.
	31	22	Do.
WBO, Asheville, N. C	33	19	Lowest for date.
	30	22	Do.

*Lowest, as used here, means lowest temperature ever recorded so late in the season; record means that the temperature equalled the previous low temperature record for so late in the season.

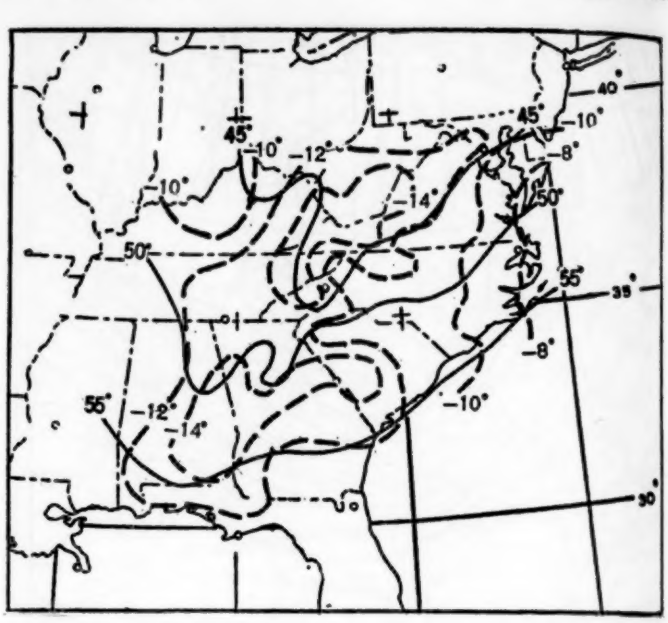

Figure 11.—Isopleths of average surface temperatures (° F.) and their departures from normal for the 4-day period April 19–22, 1953.

THE INTENSITY OF ALFA

Although Alfa was not solely responsible for the many record low temperatures that were established, on the sea level map (fig. 2 or 3) this High gave the impression of being a dominant pressure system and for a while covered an area somwehat greater than half of the United States. It might therefore be interesting to appraise the intensity of Alfa in its own right, if possible.

Intensity criteria for both Highs and Lows have been proposed by James [6] who has urged that objectivity be introduced into these evaluations. James computed the mean central pressures and standard deviations for both Highs and Lows in 5-degree latitudinal strips around the Northern Hemisphere. He then proposed, for example, that a High with a central pressure more than two σ (σ=standard deviation) higher than the mean of its corresponding latitude should be classed as very intense; a High with a central pressure greater than the mean for its latitude by an amount anywhere between σ and two σ would be rated intense, etc. But there may be a longitudinal aspect worthy of consideration in an evaluation of intensity, particularly some allowance for the climatic influences of continents, oceans, and extensive mountain areas. To illustrate this, consider latitude 32½° N., where, according to James, the mean central pressure of Highs in spring is 1,026 mb., σ^2 is 30, and a High with a central pressure of 1,040 mb., being more than 2σ greater than the mean, would be rated as very intense. Now on the basis of figure 12, showing maximum pressures for one of the spring months for the entire period of record, such a rating of very intense seems appropriate for central Texas but a value even 10 mb. lower, i. e., 1,030 mb. should deserve a very intense rating if observed in southern

Arizona. Illustrations like the above could be multiplied many fold. It therefore seems that longitudinal differences frequently deserve consideration in determinations of intensity. Until more complete studies are forthcoming, it may be possible to adjust James' ratings when they are seen to be inconsistent with extreme values based on a long period of record for the particular region.

TABLE 2.—*Intensity ratings of Alfa, following James [6]*

Latitudinal position of Alfa	Mean central pressure (mb.)	σ^2 (mb.)	Representative central pressure of Alfa (mb.)	Intensity rating
55-60° N	1,030	53	1,040	Intense.
50-55	1,029.5	37	1,037	Do.
45-50	1,028.5	44	1,037	Do.
40-45	1,027	26.5	1,033	Do.
35-40	1,025?	26	1,030	Normal.
30-35	1,026?	30	1,028	Do.
25-30	1,023	10	1,024	Do.

The intensity ratings of Alfa at the respective latitudes according to James' criteria, are listed in table 2. Alfa fell into the intense, but not the very intense, category, until it passed south of 40° N., after which its rating was of normal intensity.

Another estimate of intensity may be had from the diagrams relating frequency to central pressures of anticyclones against latitude for the Northern and Southern Hemispheres, but only for the winter and summer seasons, as published recently by Gibbs [7]. Gibbs' diagrams indicate that Alfa was somewhat less intense than 50 percent of winter anticyclones throughout its track from south central Canada through the United States, but when compared with summer anticyclones, Alfa was of extreme intensity while the center was north of the United States border, but dropped off rapidly in intensity to reach approximately a median value while moving southward through central United States.

The modification of the James technique for estimating intensity, as suggested above, is applicable to Lows as well as to Highs. This method involves the use of monthly charts of sea level pressure extremes, with isopleths drawn as shown in figure 12, obtained from Lennahan [8], along with monthly charts of sea level pressure normals, as given in [4]. The intensity of a High (or Low) would then be judged by how far its central pressure departed from the normal in the direction of the extreme, both the normal and the extreme being dependent on the longitude as well as the latitude of the High or vortex center. Values of σ which would be useful in the same manner as in the James technique, are admittedly lacking and should be incorporated into any future refinement.

In figure 13 can be seen just where the central pressure of Alfa fits in between the highest April pressures ever recorded and the April normals for each respective position along its track. From the figure, note how Alfa plots about two-thirds of the way from normal toward the extreme in the period from 1230 GMT of the 19th until 1830 GMT of the 20th, after which its central pressure

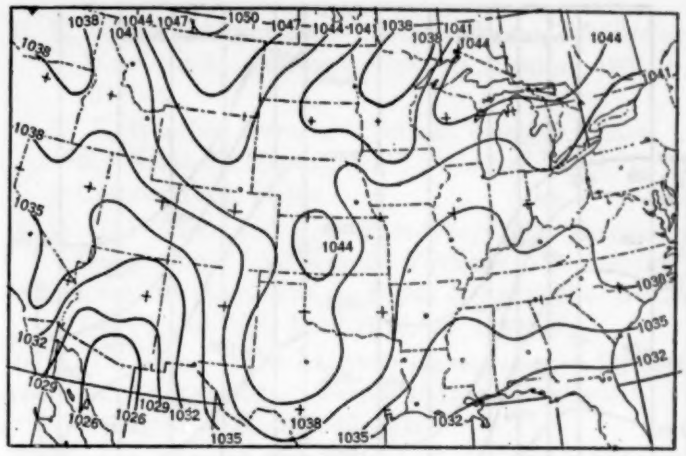

FIGURE 12.—Maximum observed sea level pressure (mb.) for April, based on entire period of record at each station. (After Lennahan [8].)

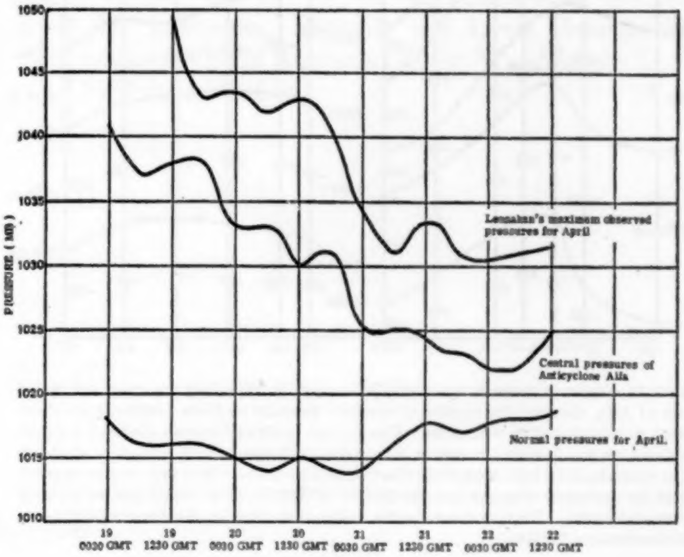

FIGURE 13.—Curves depicting normal April sea level pressure, highest pressure ever observed, and actual pressure of Alfa's center for successive 6 hourly positions along its track, with no corrections for diurnal variations.

plots closer to the normal by an amount less than half the spread between normal and extreme. Qualitatively, we conclude that Alfa was fairly intense until 1830 GMT of the 20th, after which it weakened and maintained about normal intensity. Certainly in this one case there has been close agreement concerning the intensity among the three methods.

DIFFERENTIAL TEMPERATURE CHANGES ALOFT OVER AND NEAR ALPHA

Figure 14 is essentially a space cross section showing the height profiles of selected constant pressure surfaces along the approximate major and minor axes of Alfa's sea level configuration at 0300 GMT, April 20. The center of the High at this time was about 2 degrees of longitude east of Denver, Colo. At 500 mb. Las Vegas, Nev. was

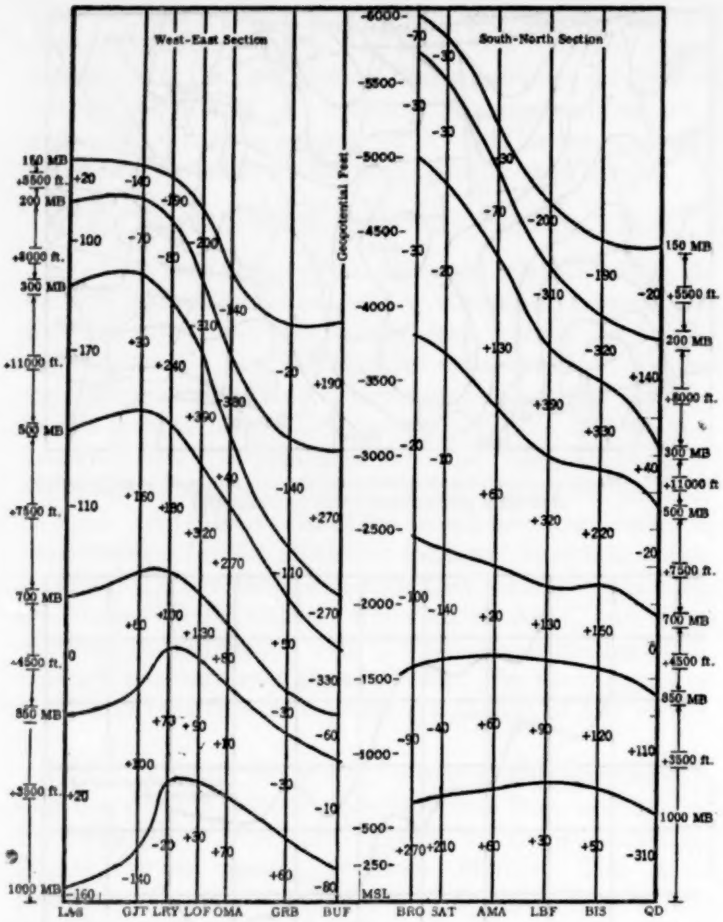

FIGURE 14.—Space cross sections at 0300 GMT, April 20, 1953 along the major and minor axes of Alfa, showing the profiles of constant pressure surfaces. Stations are those listed in table 3 and in that order. The figures midway between constant pressure surfaces are the 24-hour changes in feet of the thickness in each respective stratum. The numbers along both sides of the chart show the number of feet that were subtracted from the particular constant pressure profiles so that the chart would not be too large for reproduction. These numbers are the values that must be added to obtain the true thicknesses and heights.

just east of a cold Low along the Pacific Coast, and Green Bay, Wis. and Buffalo, N. Y. were under the influence of a cold Low centered near Buffalo. Brownsville, Tex. was in the direction toward which Alfa was moving while The Pas, Manitoba was in the area from which Alfa was rapidly departing.

In the chart, which extends to 150 mb., thicknesses between the various constant pressure surfaces (also sea level to 850 mb.) have been depressed by variable amounts as found to be expedient. The number of feet that should be added to the heights of the constant pressure surfaces, is indicated along the ordinate scale on both edges of the chart. If the actual height of, say, the 300 mb. level at Grand Junction, Colo. is wanted, it can be obtained by taking 4,250 feet read from the center ordinate scale, and adding 3,500 feet (depression of MSL–850 mb. thickness) plus 4,500 feet (850–700 mb. depression) and correspondingly, plus 7,500 feet, plus 11,000 feet to give a total of 30,750 feet. While the various strata have not been uniformly depressed, the differences in thickness for a particular stratum between stations are exact. Thus just a glance at the chart indicates that through the western section, Las Vegas to Denver, all strata from 1,000 mb. to 300 mb. were thicker than those through the eastern section, Denver to Buffalo. Contrariwise, above 300 mb. the strata between constant pressure surfaces were thinner in the west, thicker in the east.

TABLE 3.—*24-hour height changes of selected constant pressure surfaces and in thickness between selected constant pressure levels (assuming a constant height for the "0"-mb. level) at stations along the major and minor axes of Alfa during period ending at 0300 GMT, Apr. 20, 1953*

West-East Section

24-hour change in—	Las Vegas, Nev.	Grand Junction, Colo.	Denver, Colo.	North Platte, Nebr.	Omaha, Nebr.	Green Bay, Wis.	Buffalo, N. Y.
1,000-mb. height	−160	−140	−20	+30	+70	+60	−80
150-mb. height	−500	0	+300	+450	−50	−220	−290
1,000-150 mb. thickness	−340	+140	+320	+420	−120	−280	−210
150-0 mb. thickness	+500	0	−300	−450	+50	+220	+290

South-North Section

24-hour change in—	Brownsville, Tex.	San Antonio, Tex.	Amarillo, Tex.	North Platte, Nebr.	Bismarck, N. Dak.	The Pas, Manitoba
1,000-mb. height	+270	+210	+60	+30	+50	−310
150-mb. height	−70	−60	+230	+450	+360	−60
1,000-150 mb. thickness	−340	−270	+170	+420	+310	+250
150-0 mb. thickness	+70	+60	−230	−450	−360	+60

Along the west-east section Alfa was warmer in the west than in the east below 300 mb. and the reverse was true above 300 mb. In the south-north section some of the slope of the constant pressure surfaces downward to the north may be a normal latitudinal influence. Note how the mean temperatures associated with the various strata were generally higher in the south than in the north below 300 mb. while it was perceptibly colder in the southern section than in the northern in the 200–150 mb. stratum. More specifically, the 200–150 mb. thickness at Brownsville was 5,730 feet, equivalent to a mean virtual temperature of about −65.6° C., and at The Pas, it was 6,150 feet, which is equal to about −50.4° C.

Since changes in the 1,000-mb. height are very closely correlated with sea level pressure changes, it may be of interest to compare the 24-hour height changes of the 1,000-mb. level with the 24-hour net changes in thickness of the 1,000–150 mb. stratum for the stations in the cross section. Some of the computations are tabulated in table 3. Let us assume a constant height for the "zero"-mb. level and make a sample deduction. At Las Vegas the 1,000-mb. height fell 160 feet and the 150-mb. height fell 500 feet giving a net change in the 1,000–150 mb. thickness of −340 feet. Therefore, cooling below 150 mb. was more than compensated for by warming above 150 mb. to account for the decrease in sea level pressure. If now we examine the 24-hour changes in thickness near the center

of Alfa at Denver and North Platte (fig. 14) we find that while warming had occurred below 300 mb., nevertheless sufficient cooling took place above 300 mb. to permit a small rise in the 1,000-mb. height at North Platte and only a very small fall at Denver.

Although this study has revealed some interesting features in the history, structure, and environmental interaction of the anticyclone Alfa, it has left unanswered many questions on the formation and evolution of the High. The authors conclude with Wexler [3] that "it is quite impossible to discuss the anticyclone as a separate entity, with respect to either its origin or its role in the general circulation".

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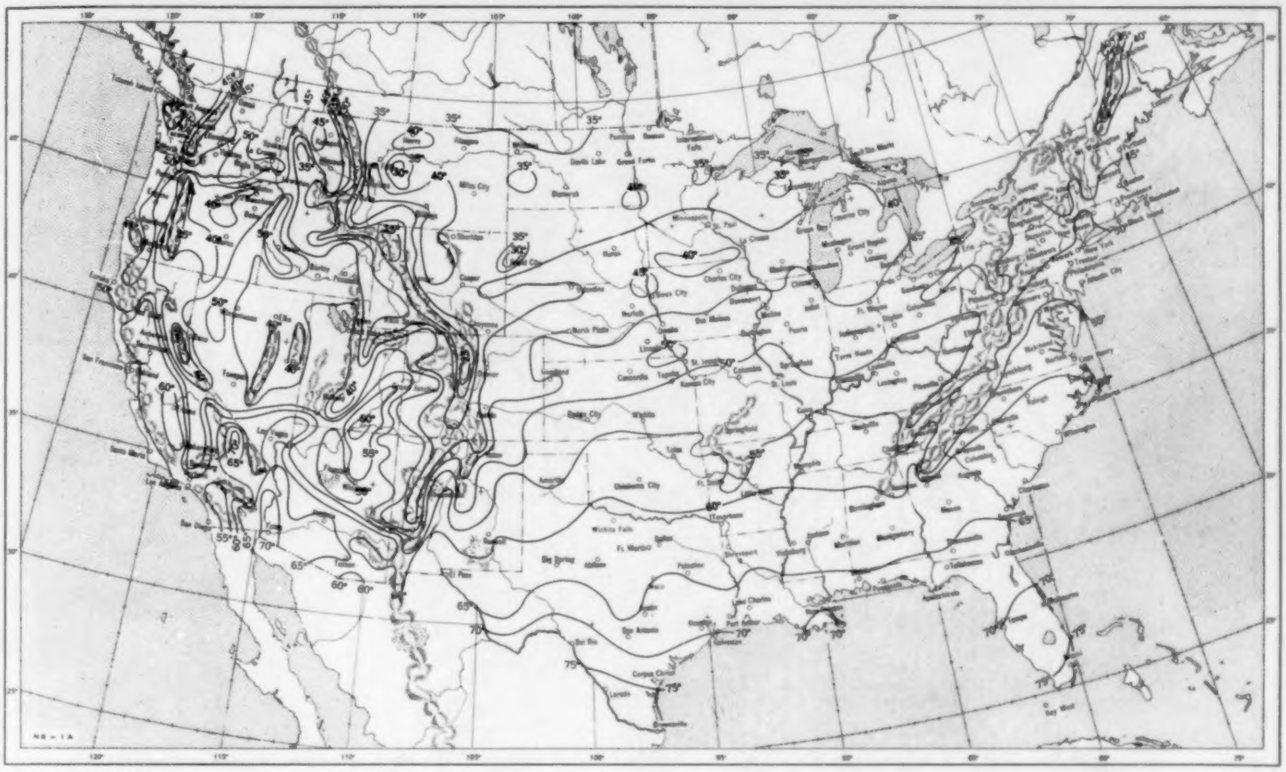
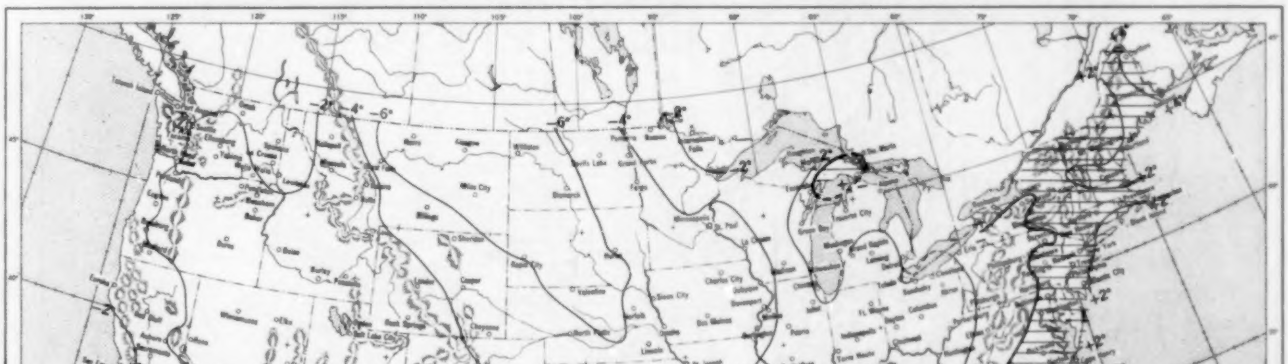
CORRECTION

MONTHLY WEATHER REVIEW, vol. 80, No. 3, March 1953, page 82: In column 2, in text beneath table 1, total March 1953 precipitation at Boston should be 11.00 in. instead of 11.69 as given. The next sentence should read "The 24-hour total of 3.10 in. for the period ending about 1900 EST, March 30 is 0.06 in. greater than any previous 24-hour total in March."

Chart I. A. Average Temperature (°F.) at Surface, April 1953.

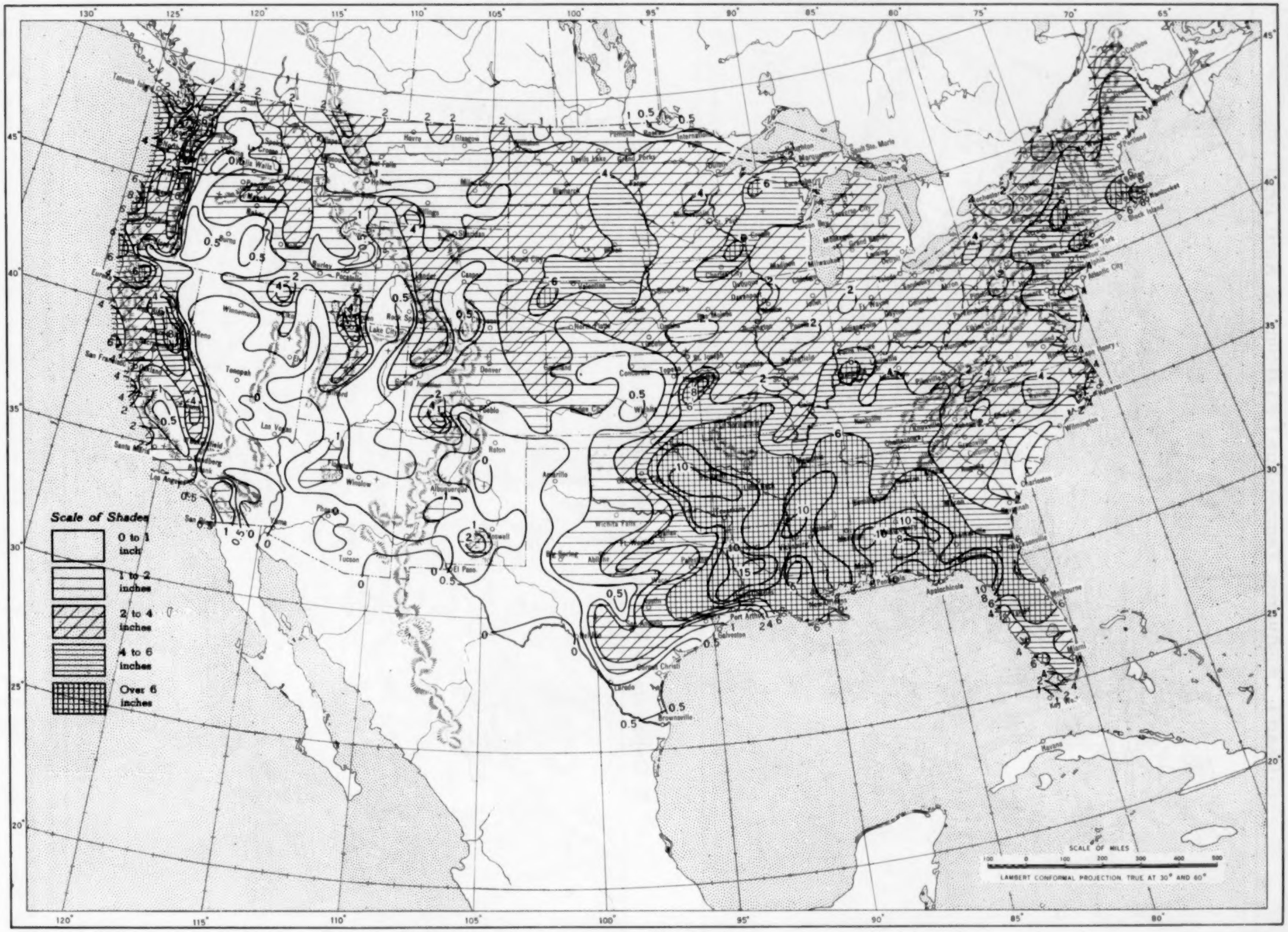


B. Departure of Average Temperature from Normal (°F.), April 1953.



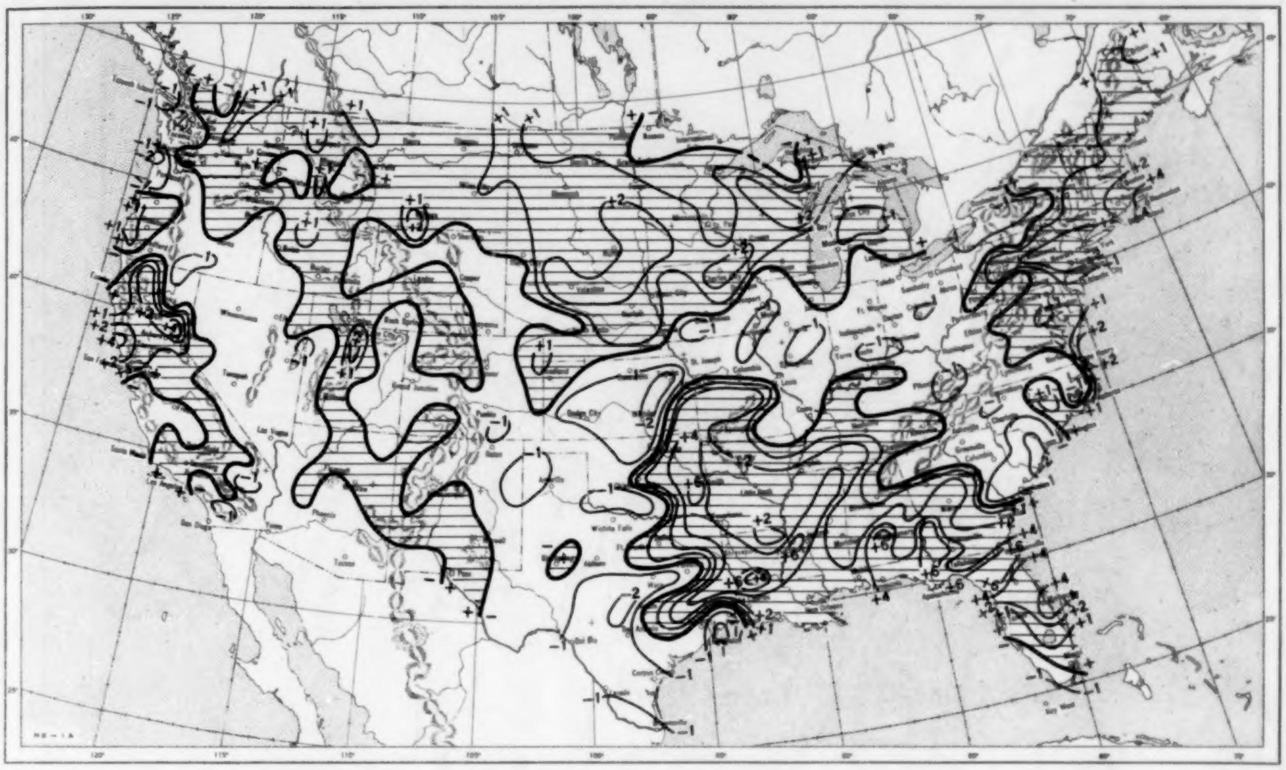
A. Based on reports from 800 Weather Bureau and cooperative stations. The monthly average is half the sum of the monthly average maximum and monthly average minimum, which are the average of the daily maxima and daily minima, respectively.
B. Normal average monthly temperatures are computed for Weather Bureau stations having at least 10 years of record.

Chart II. Total Precipitation (Inches), April 1953.

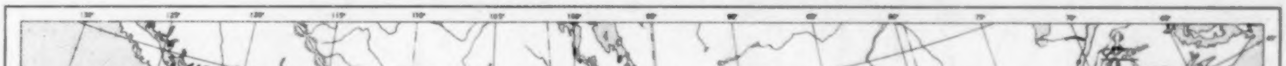


Based on daily precipitation records at 800 Weather Bureau and cooperative stations.

Chart III. A. Departure of Precipitation from Normal (Inches), April 1953.

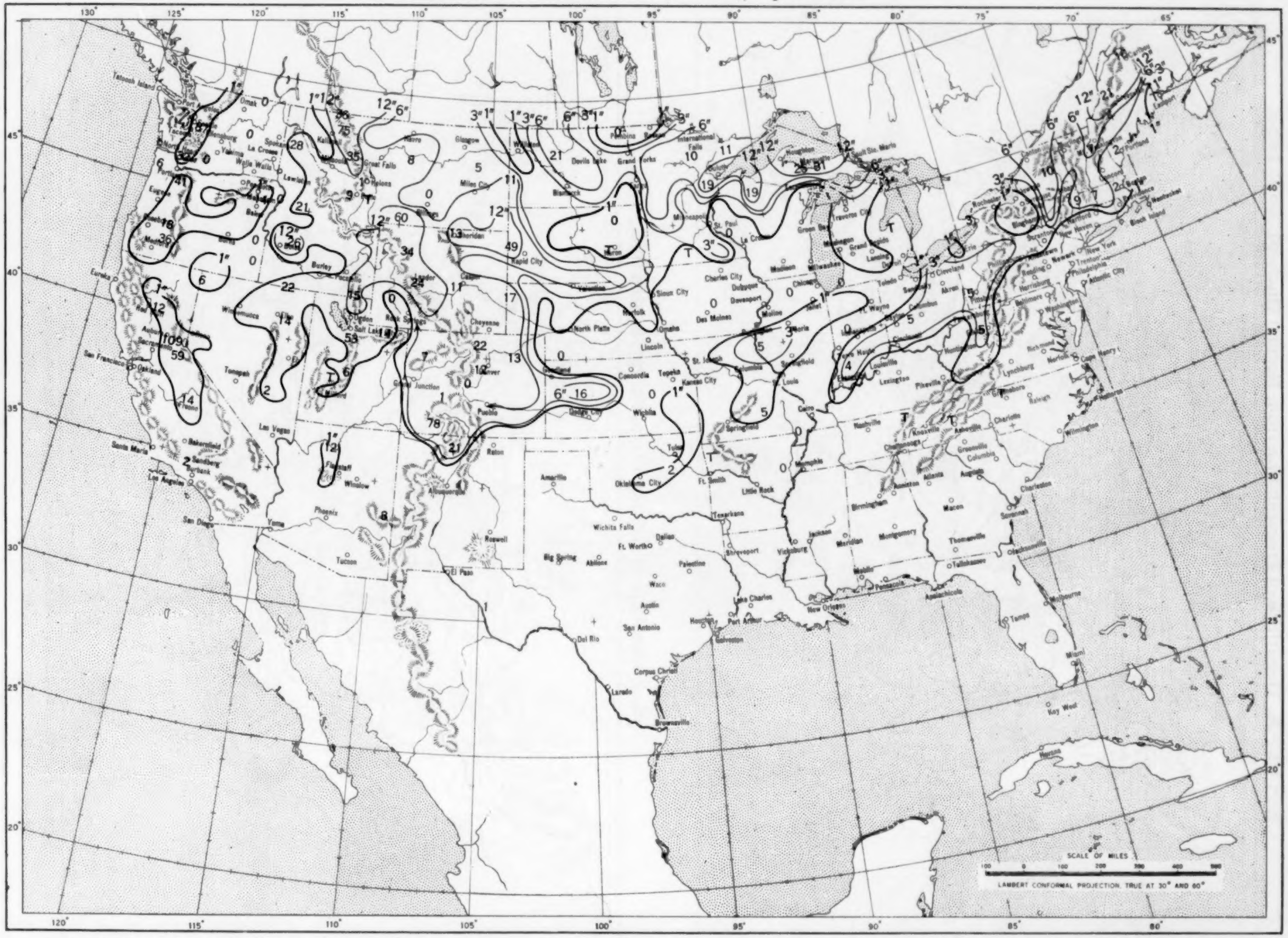


B. Percentage of Normal Precipitation, April 1953.



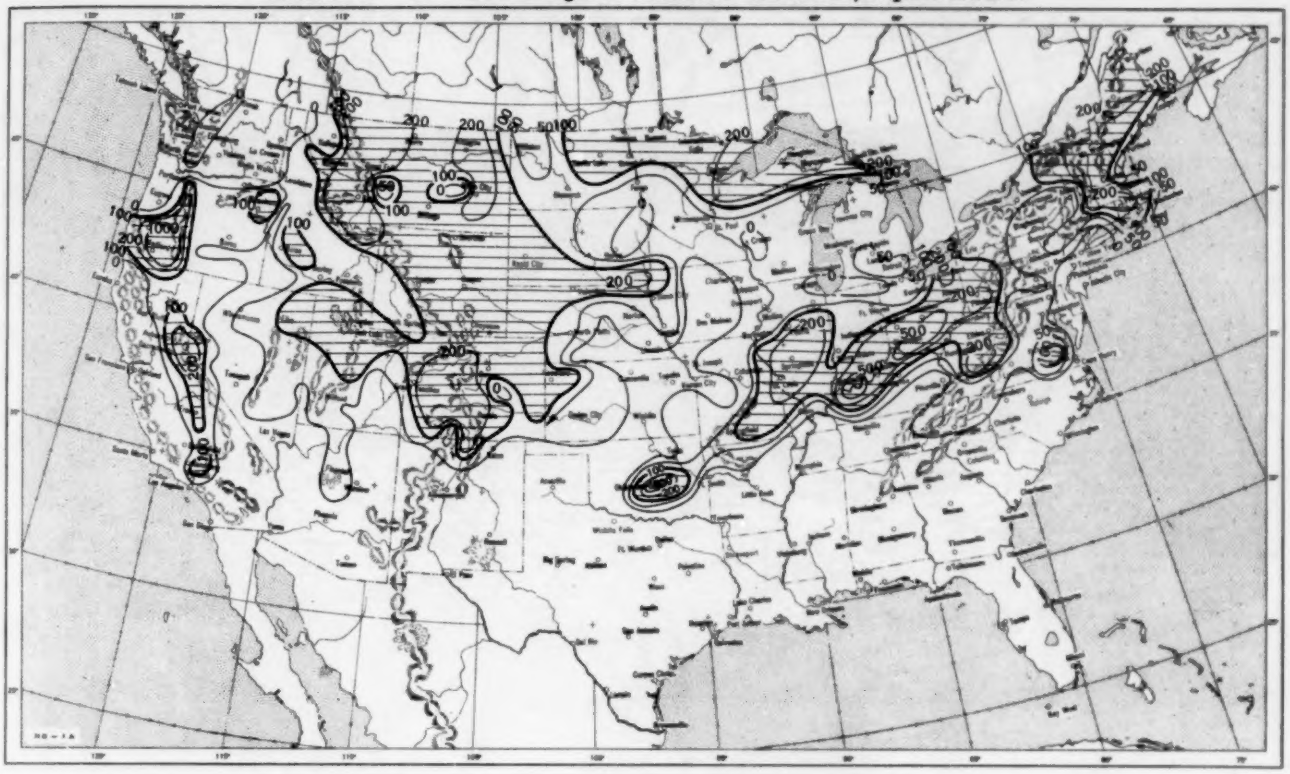
Normal monthly precipitation amounts are computed for stations having at least 10 years of record.

Chart IV. Total Snowfall (Inches), April 1953.

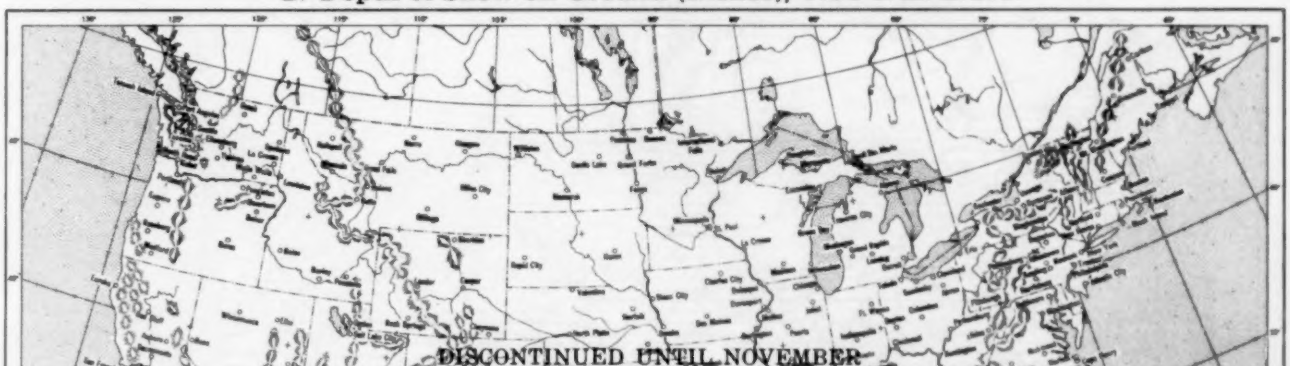


This is the total of unmelted snowfall recorded during the month at Weather Bureau and cooperative stations. This chart and Chart V are published only for the months of November through April although of course there is some snow at higher elevations, particularly in the far West, earlier and later in the year.

Chart V. A. Percentage of Normal Snowfall, April 1953.

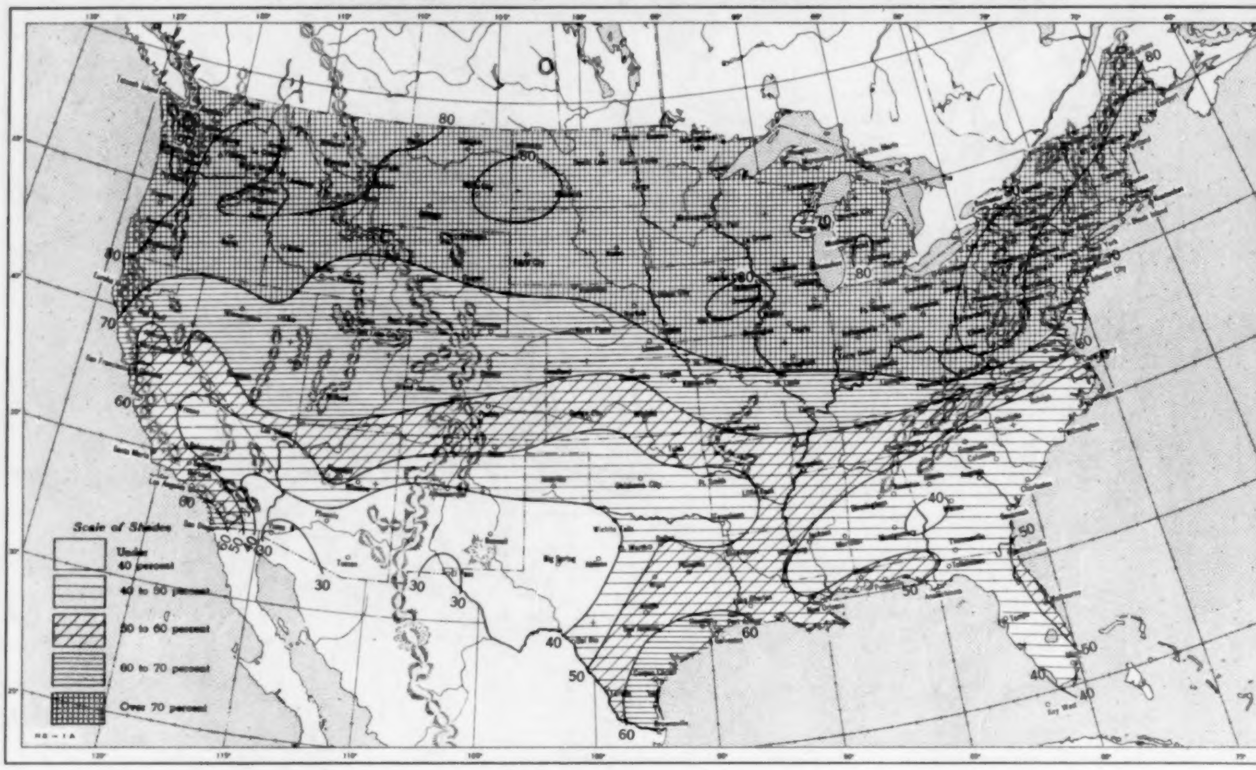


B. Depth of Snow on Ground (Inches), 7:30 a. m. E. S. T.

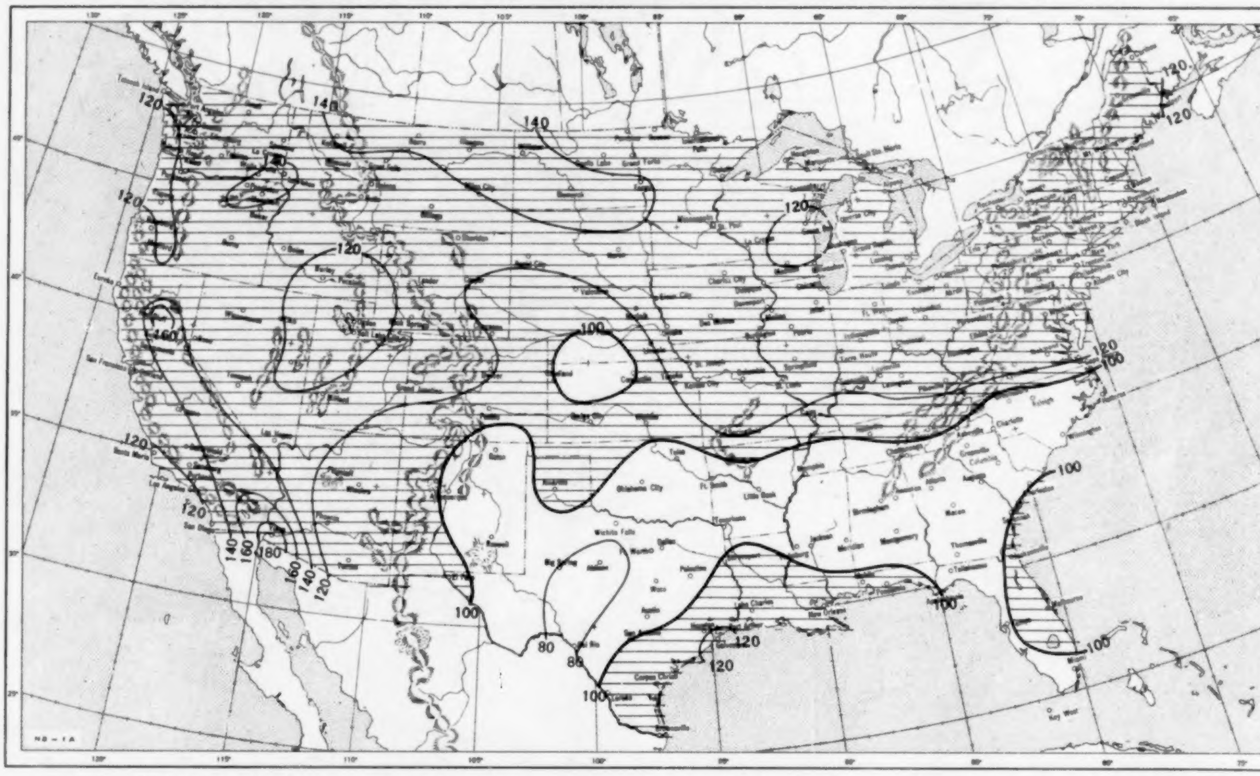


A. Amount of normal monthly snowfall is computed for Weather Bureau stations having at least 10 years of record.
B. Shows depth currently on ground at 7:30 a. m. E. S. T., of the Tuesday nearest the end of the month. It is based on reports from Weather Bureau and cooperative stations. Dashed line shows greatest southern extent of snowcover during month.

Chart VI. A. Percentage of Sky Cover Between Sunrise and Sunset, April 1953.

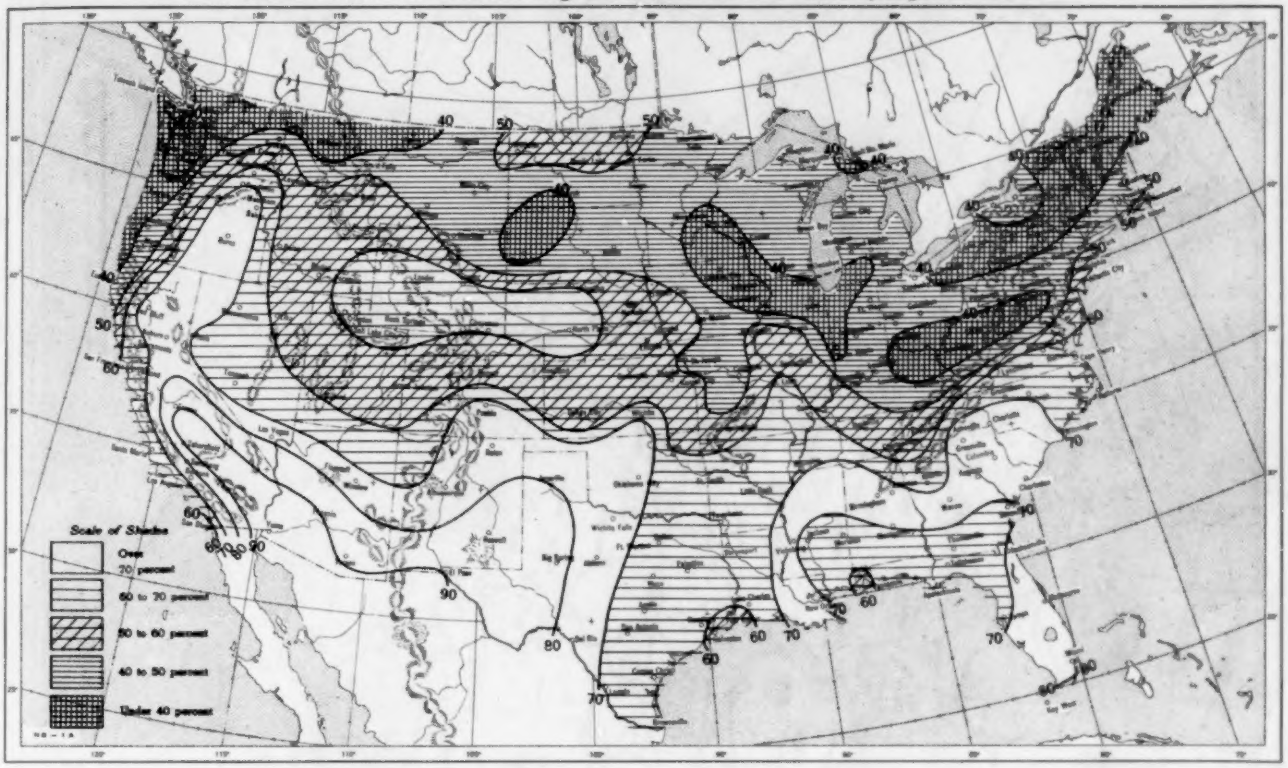


B. Percentage of Normal Sky Cover Between Sunrise and Sunset, April 1953.

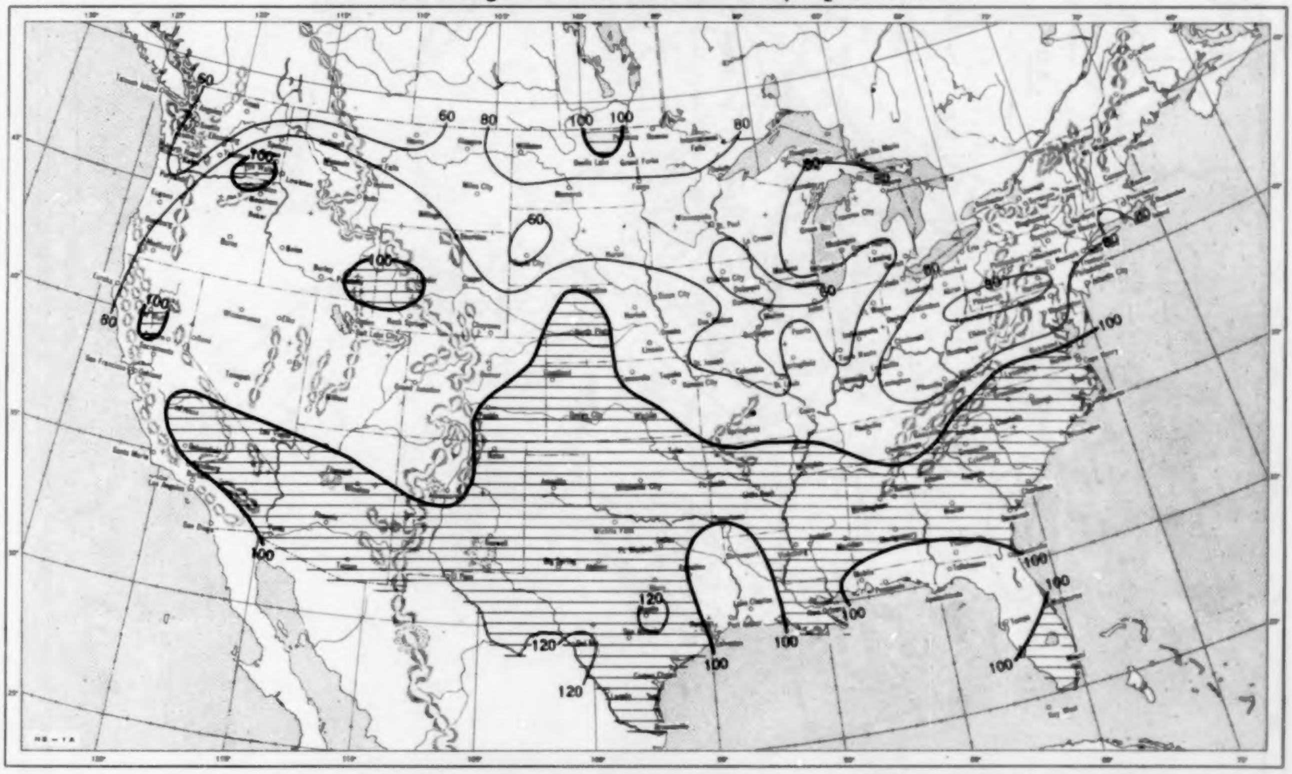


A. In addition to cloudiness, sky cover includes obscuration of the sky by fog, smoke, snow, etc. Chart based on visual observations made hourly at Weather Bureau stations and averaged over the month. B. Computations of normal amount of sky cover are made for stations having at least 10 years of record.

Chart VII. A. Percentage of Possible Sunshine, April 1953.



B. Percentage of Normal Sunshine, April 1953.



A. Computed from total number of hours of observed sunshine in relation to total number of possible hours of sunshine during month. B. Normals are computed for stations having at least 10 years of record.

Chart VIII. Average Daily Values of Solar Radiation, Direct + Diffuse, April 1953. Inset: Percentage of Normal Average Daily Solar Radiation, April 1953.

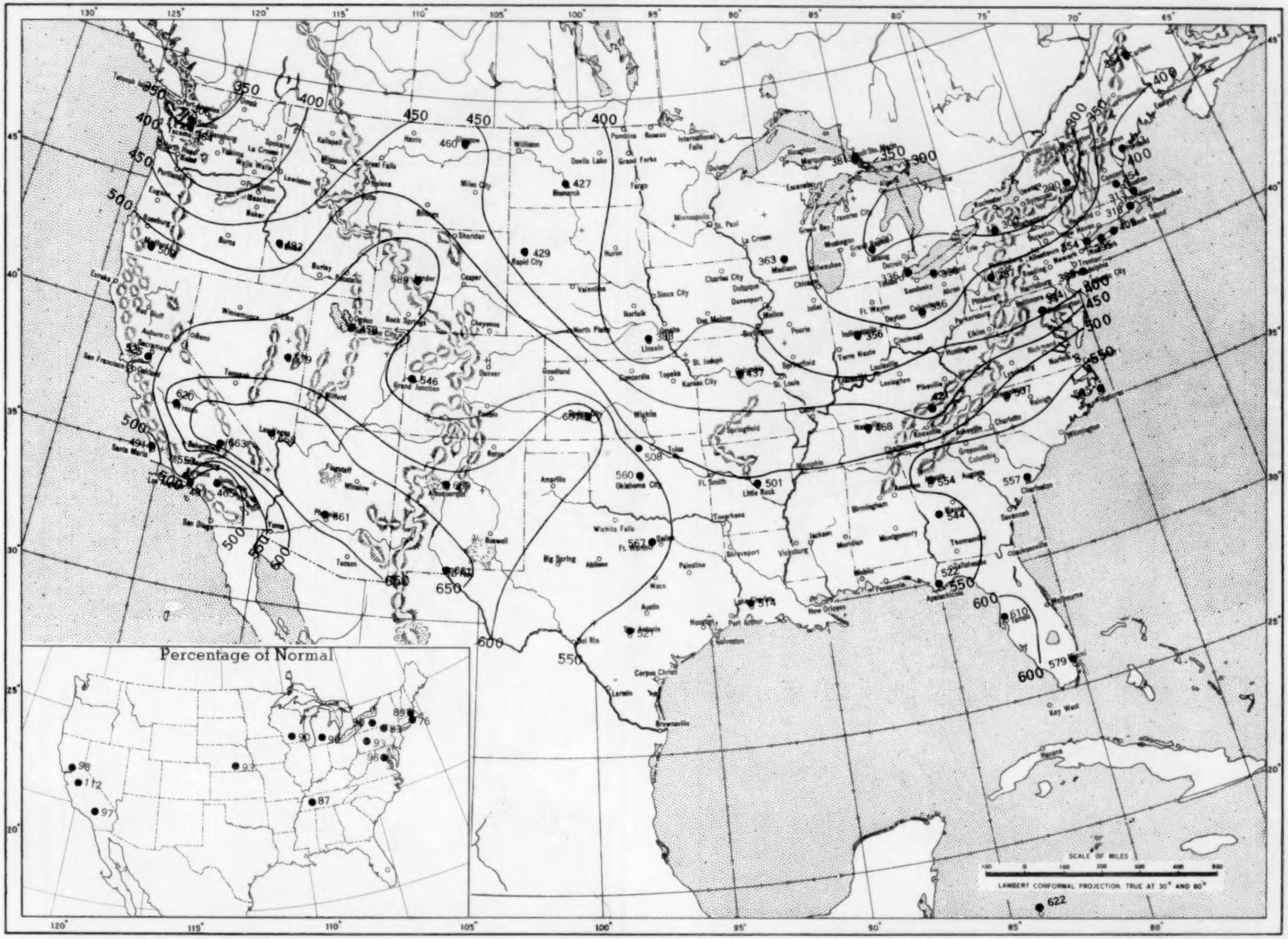
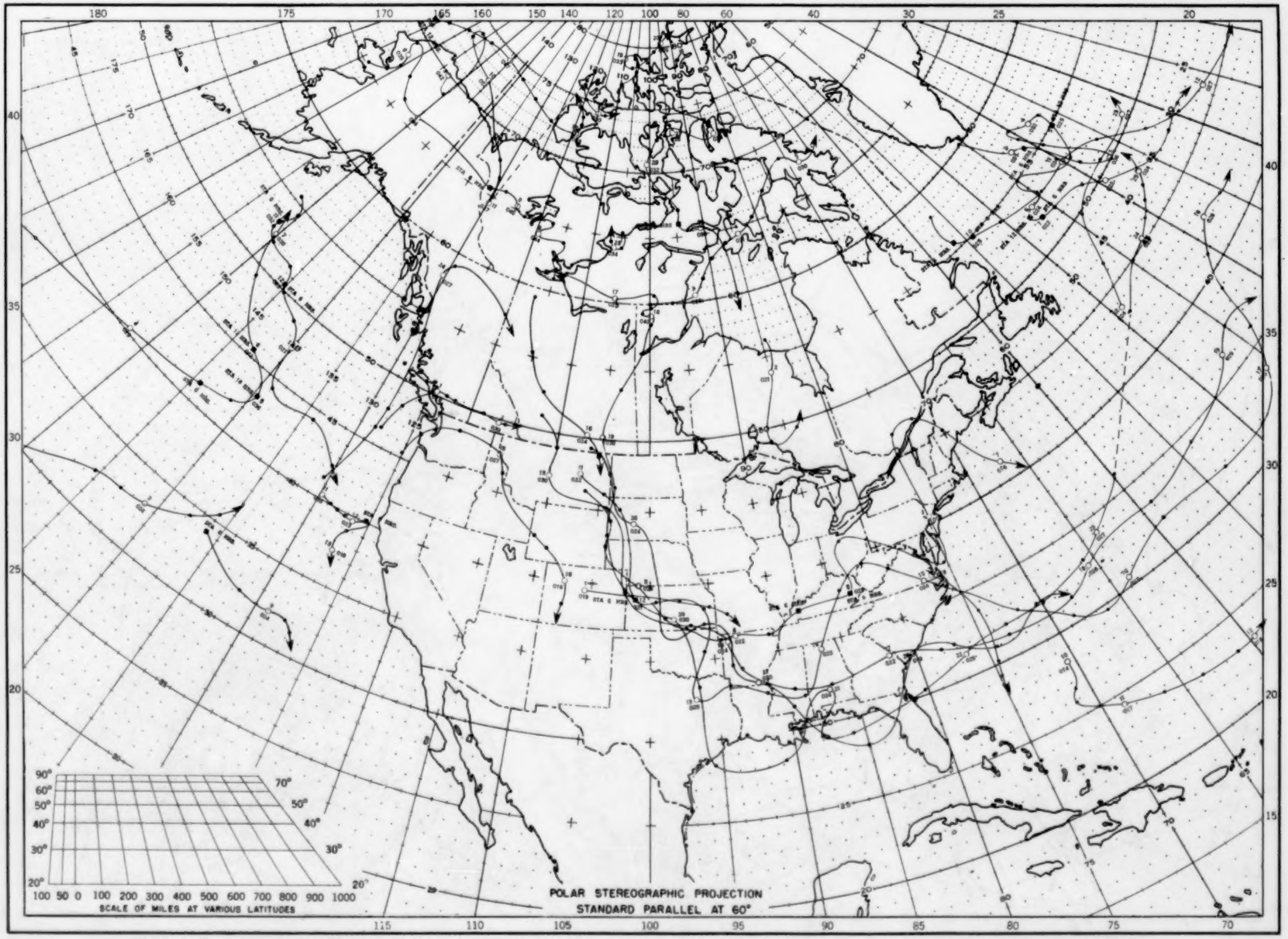


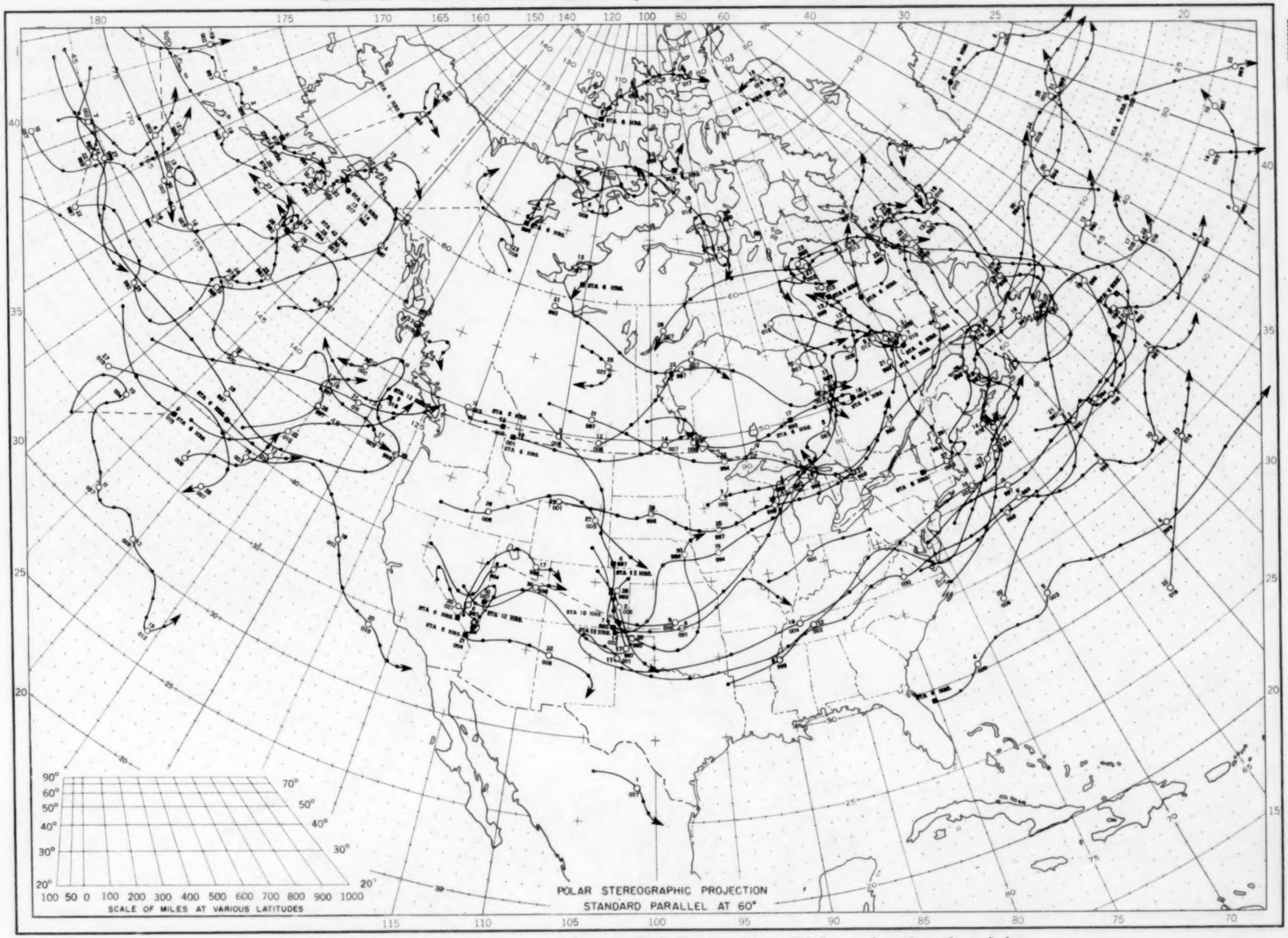
Chart shows mean daily solar radiation, direct + diffuse, received on a horizontal surface in langleys (1 langley = 1 gm. cal. cm.$^{-2}$). Basic data for isolines are shown on chart. Further estimates are obtained from supplementary data for which limits of accuracy are wider than for those data shown. Normals are computed for stations having at least 9 years of record.

Chart IX. Tracks of Centers of Anticyclones at Sea Level, April 1953.



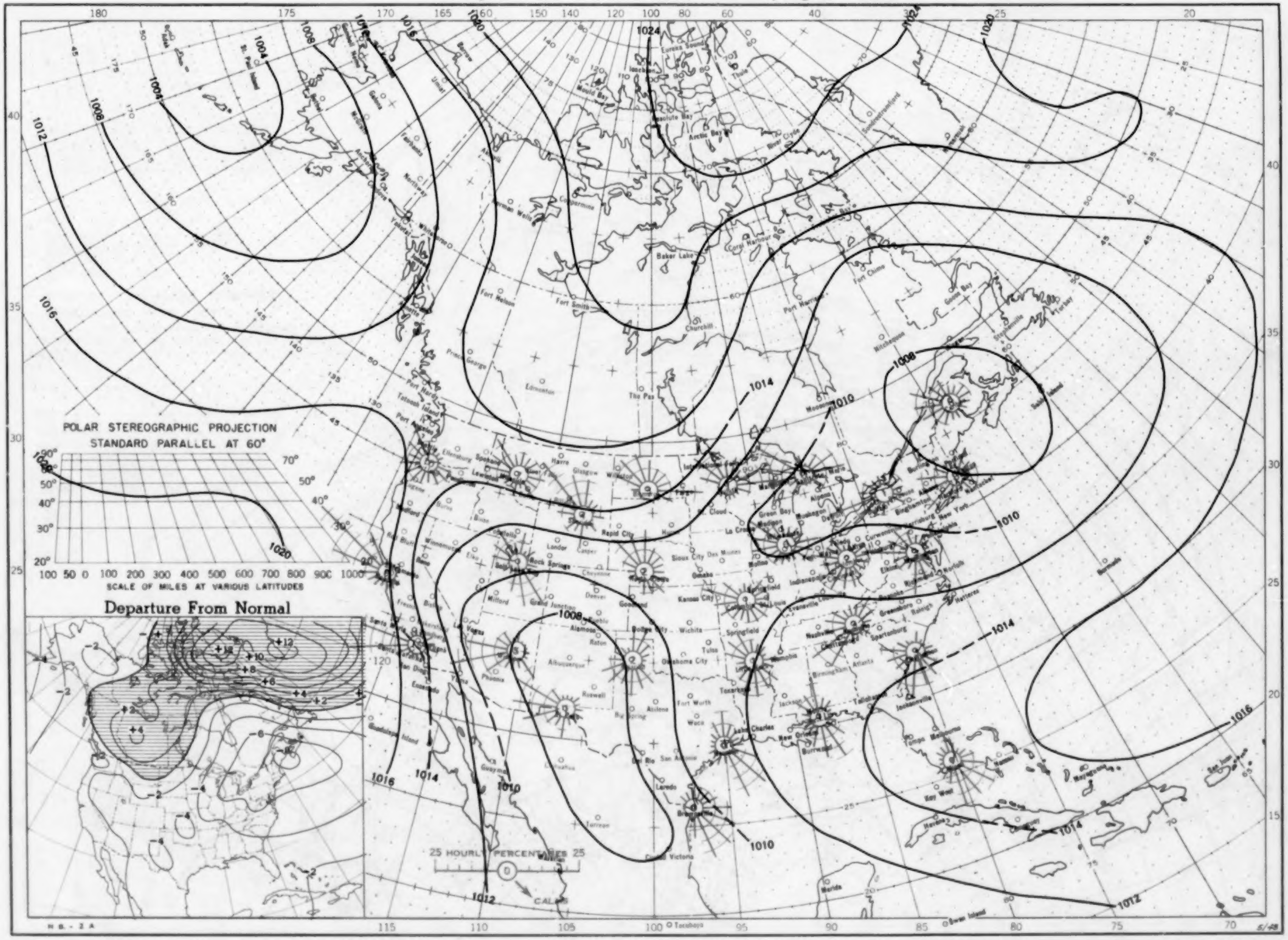
Circle indicates position of center at 7:30 a. m. E. S. T. Figure above circle indicates date, figure below, pressure to nearest millibar. Dots indicate intervening 6-hourly positions. Squares indicate position of stationary center for period shown. Dashed line in track indicates reformation at new position. Only those centers which could be identified for 24 hours or more are included.

Chart X. Tracks of Centers of Cyclones at Sea Level, April 1953.



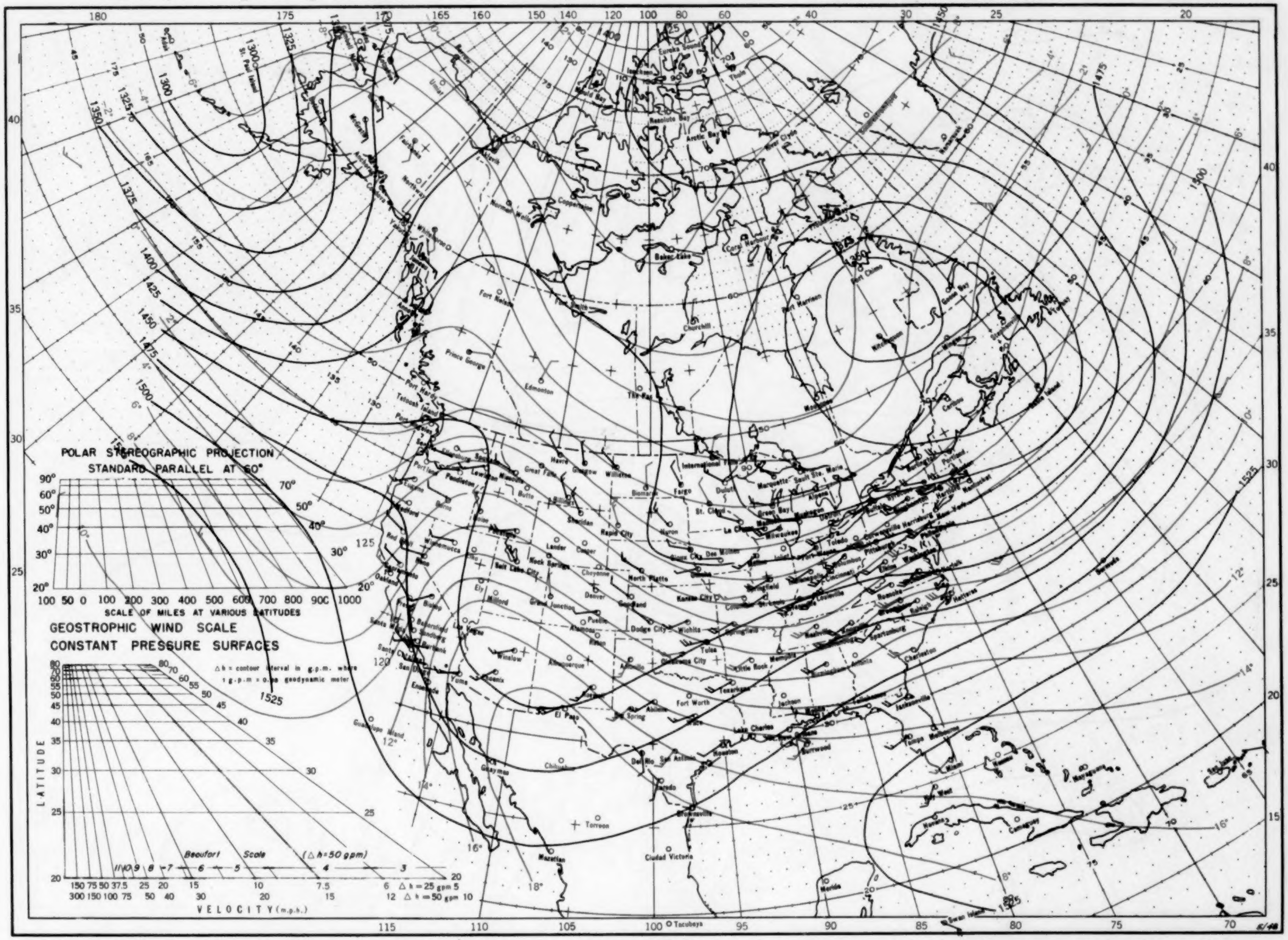
Circle indicates position of center at 7:30 a. m. E. S. T. See Chart IX for explanation of symbols.

Chart XI. Average Sea Level Pressure (mb.) and Surface Windroses, April 1953. Inset: Departure of Average Pressure (mb.) from Normal, April 1953.



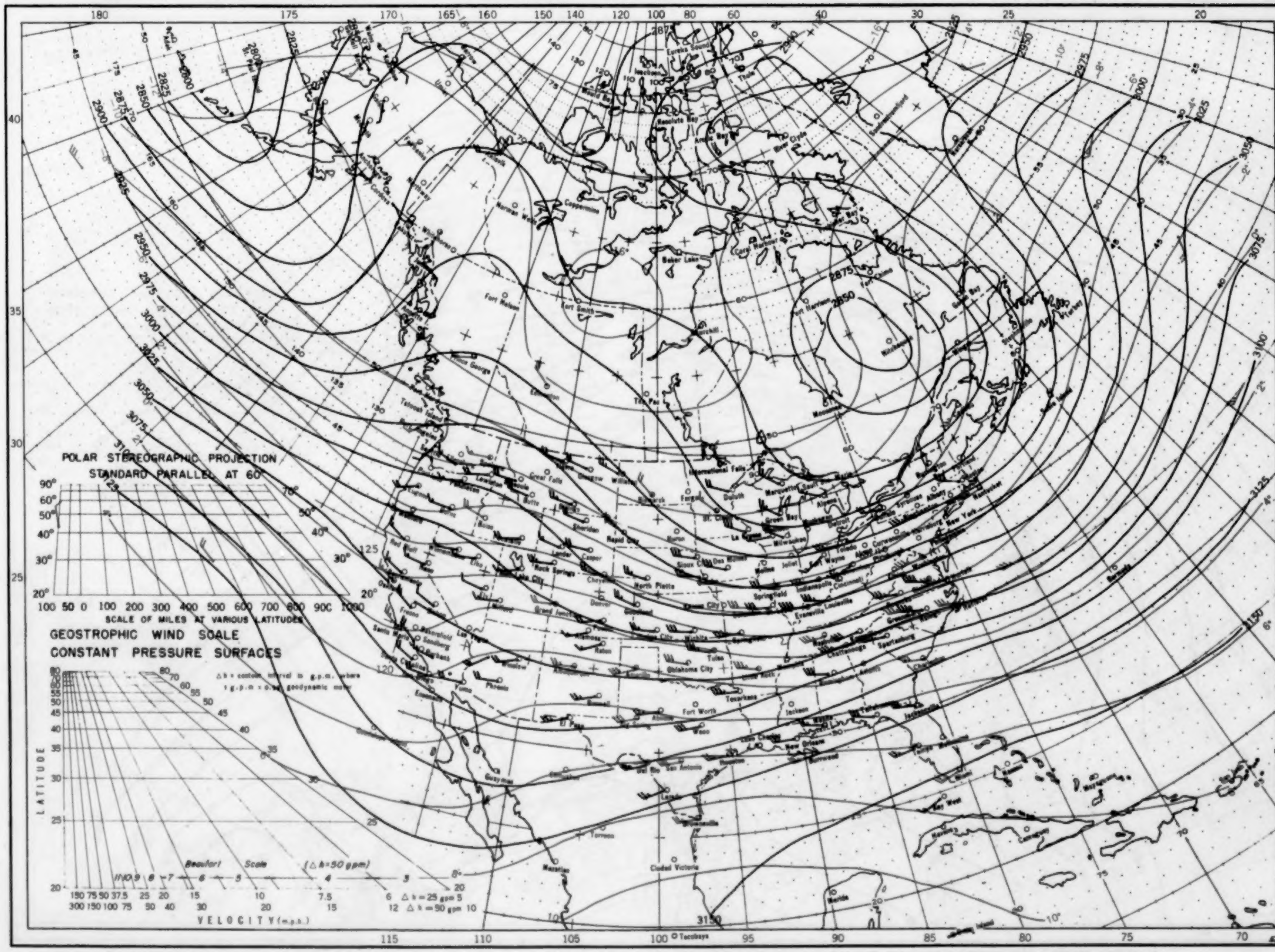
Average sea level pressures are obtained from the averages of the 7:30 a. m. and 7:30 p. m. E. S. T. readings. Windroses show percentage of time wind blew from 16 compass points or was calm during the month. Pressure normals are computed for stations having at least 10 years of record and for 10° intersections in a diamond grid based on readings from the Historical Weather Maps (1899-1939) for the 20 years of most complete data coverage prior to 1940.

Chart XII. Average Dynamic Height in Geopotential Meters (1 g.p.m. = 0.98 dynamic meters) of the 850-mb. Pressure Surface, Average Temperature in °C. at 850 mb., and Resultant Winds at 1500 Meters (m.s.l.), April 1953.



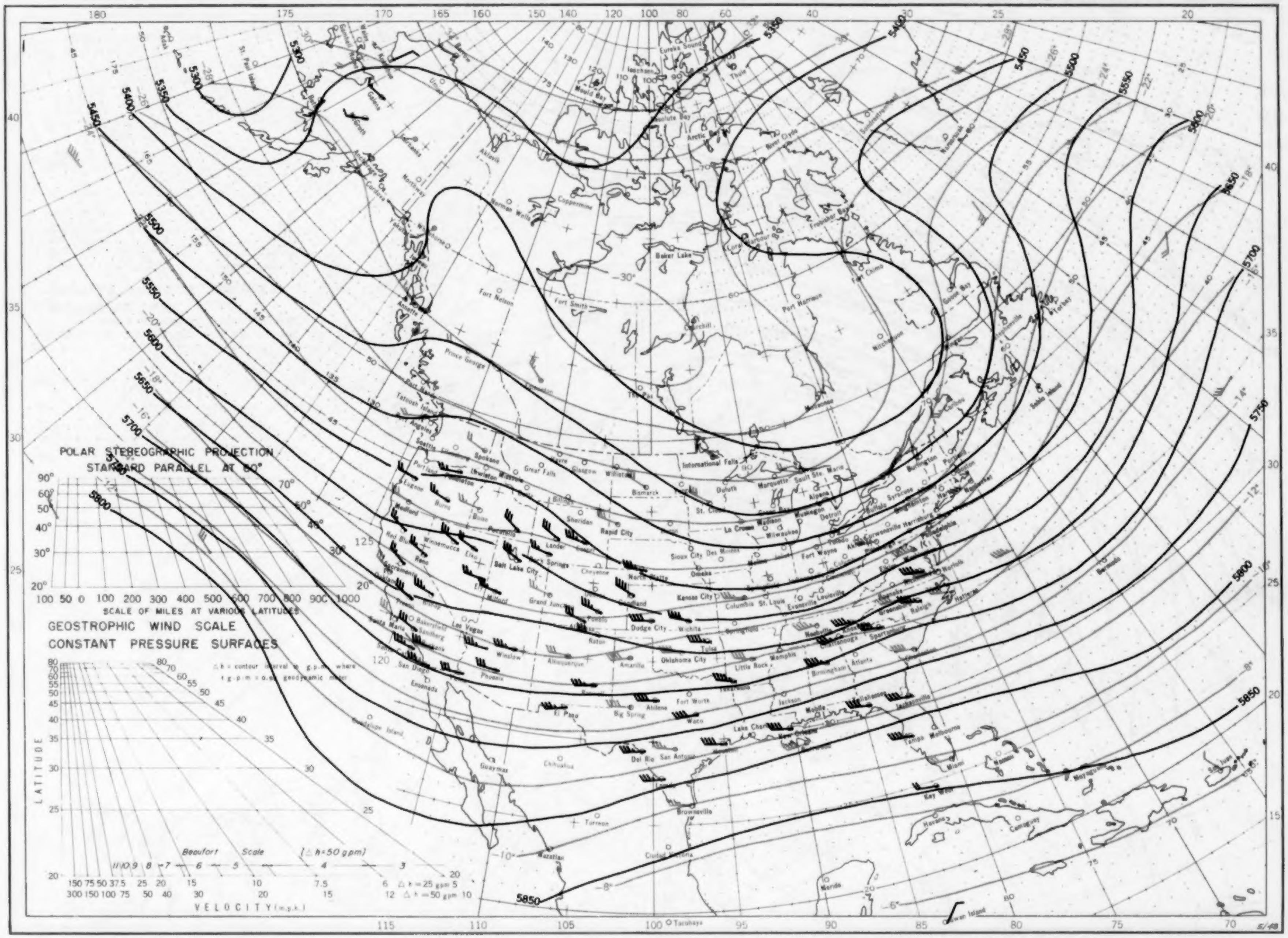
Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins taken at 0300 G. M. T.

Chart XIII. Average Dynamic Height in Geopotential Meters (1 g.p.m.. = 0.98 dynamic meters) of the 700-mb. Pressure Surface, Average Temperature in °C. at 700 mb., and Resultant Winds at 3000 Meters (m.s.l.), April 1953.



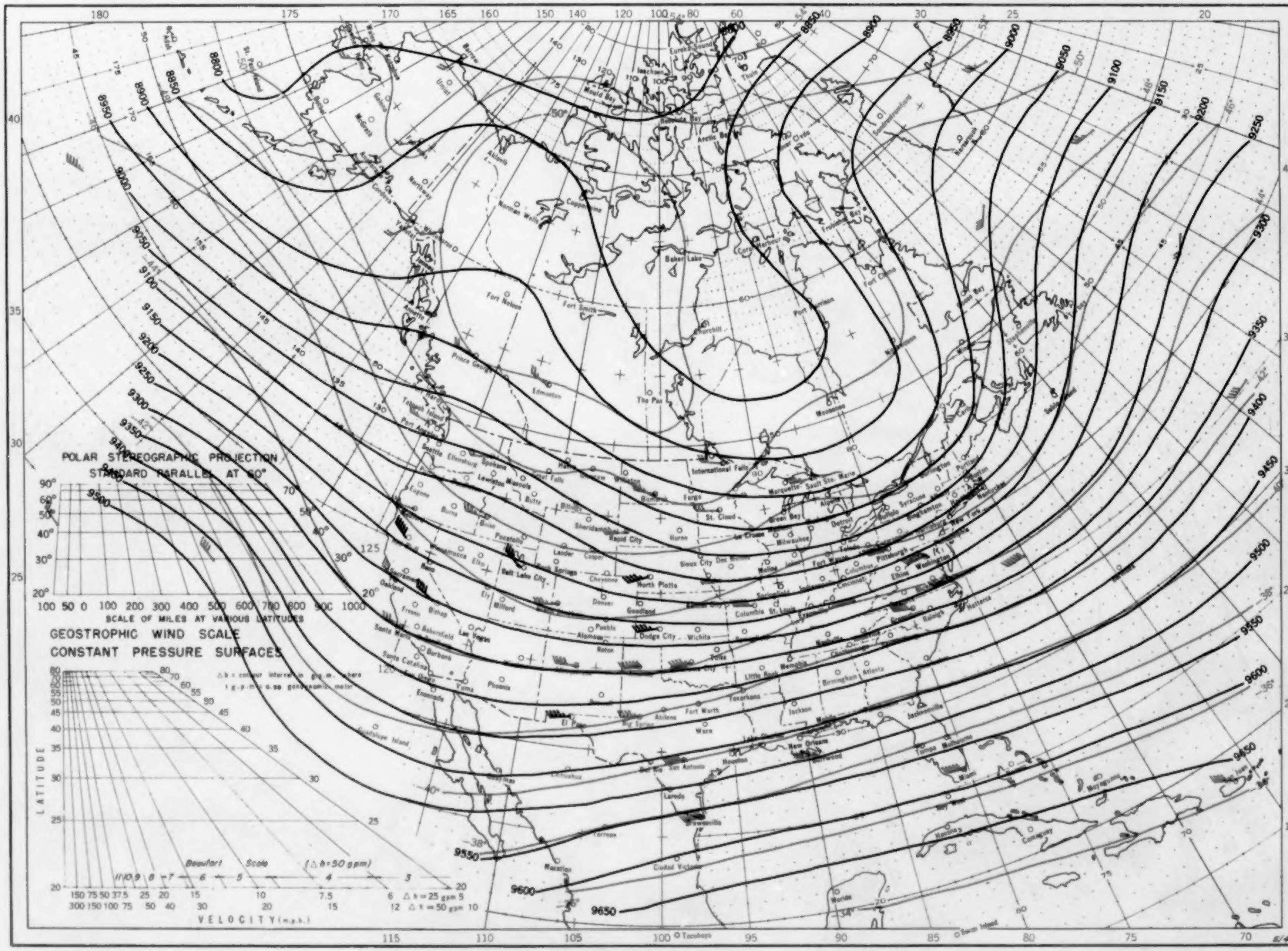
Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins taken at 0300 G. M. T.

Chart XIV. Average Dynamic Height in Geopotential Meters (1 g.p.m. = 0.98 dynamic meters) of the 500-mb. Pressure Surface, Average Temperature in °C. at 500 mb., and Resultant Winds at 5000 Meters (m.s.l.), April 1953.



Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins at 0300 G. M. T.

Chart XV. Average Dynamic Height in Geopotential Meters (1 g.p.m. = 0.98 dynamic meters) of the 300-mb. Pressure Surface, Average Temperature in °C. at 300 mb., and Resultant Winds at 10,000 Meters (m.s.l.), April 1953.



Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins at 0300 G. M. T.